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BULLETIN
OF
The Geological Society of America

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VOLUME 60 NUMBER 11

NOVEMBER 1949

PUBLISHED MONTHLY BY THE SOCIETY

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Bulletin
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VOLUME 60 November 1949 NUMBER 11

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Subscription $10 per year.

Publication Office: Mt. Royal & Guilford Aves., Baltimore 2, Md.

Communications for publication should be addressed to The Geological Society of America, Dr. H. R. Aldrich, Secretary, 419 West 117th Street, New York 27, N. Y.

NOTICE.—Four weeks notice of change of address is necessary. In accordance with the rules established by the Council, claims for non-receipt of the preceding number of the Bulletin must be sent to the Secretary of the Society within three months of the date of the receipt of this number in order to be filled gratis.

Entered as second-class matter in the Post-Office at Baltimore, Md., under the Act of Congress of July 16, 1894.

Accepted for mailing at special rate of postage provided for in Section 1103 Act of October 3, 1917, authorized on July 8, 1918.

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BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 60, PP. 1731-1754, 2 PLS., 11 FIGS. NOVEMBER 1949

RECONNAISSANCE GEOLOGY AND VERTEBRATE PALEONTOLOGY OF EL SALVADOR, CENTRAL AMERICA

By R. A. Stirton and W. K. Gealey

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Abstract

The surface rocks in El Slavador consist primarily of ?Pliocene, early Pleistocene, late Pleistocene, and recent volcanics.

In the Metapán area are exposures of ?pre-Mesozoic monzonite, the lower and middle Jurassic continental Metapán beds, and Cretaceous marine limestone, overlain unconformably by late Tertiary volcanics and continental sediment. The Tertiary is represented by a poorly consolidated sandstone and conglomerate approximately 650 feet thick, which overlies with unconformity 30 feet of gray andesite. The Mesozoic includes 400 feet of early Cretaceous limestone, underlain with apparent conformity by the Metapán beds, consisting of 180 feet of red and brown calcareous shale, 30 feet of hard vein quartz pebble conglomerate, and 75+ feet of interbedded fine-grained green sandstone and red, brown, and green shales.

No identifiable vertebrate remains older than Pleistocene were found in El Salvador, though some of the beds referred to ?Pliocene or early Pleistocene may eventually yield late Pliocene fossils. Fossils have been found in 20 localities in 8 of the 14 departments in the country. The best locality was the Hormiguero quarry where a large series of *Megatherium* remains were uncovered. These ground

sloths had been trapped and mired in a boghole. Other mammals found in the quarry were *Smilodon*, *Bison*, toxodont, and a small camelid. *Mammuthus* (mammoth) and *Cuvieronius* (spiral-tusked mastodon) were found at other localities. This probably is near the southernmost range of the mammoth and possibly near the northern limit for the South American toxodonts. Some poorly preserved fish and leaves, probably late Pleistocene, were found in barranca El Sisimico near San Vicente.

Introduction and Acknowledgments

The geological work upon which this report is based was undertaken from December, 1941, to May, 1942, in conjunction with a University of California biological expedition to El Salvador, Central America. The intention here is to present a somewhat more detailed account of the geology and the vertebrate paleontology of El Salvador than has been done previously.

Due to the very limited time available for field work, a rapid coverage of the country as a whole was considered of greater value than local detail. Hence this is a reconnaissance report, emphasizing the general distribution of rock types and their topographic expression. Lack of any large-scale map and relatively poor communications were serious handicaps in geological work.

We wish to acknowledge a generous grant from The Geological Society of America that enabled us to carry out our geological and paleontological projects. The government of El Salvador and its residents co-operated in every manner possible. We wish to express our gratitude particularly for the personal interest and support given by Mr. G. A. Swanquist, of San Miguel, and by Dr. Mario Lewy van Severen, of San Salvador. The following residents greatly facilitated field work: Sr. A. Alfaro, San Salvador; General José Tomás Calderón, San Salvador; Sr. Helio Cañas, San Miguel; Colonel Francisco Carpintero, Morazan; Sr. José Fonseca, San Salvador; Srs. Florentin and Francisco Fuentes, San Miguel; Sr. Alejandro Ghiorsi, San Salvador; Sr. Max Haltmayer, San Miguel; Sr. Pedro Mata, Los Esesmiles; Sr. Julio Media, San Salvador; Sr. Santiago Diaz Palacio, Santa Ana; Sr. Fernando Portillo, San Miguel; Sr. Enrique Prunera, San Miguel; Sr. Herman Schlageter, San Miguel; Mr. E. P. Thompson, Montecristo; Sr. Francisco Valdez, Olomega; Sr. C. K. Vilanova, San Salvador; and many others. Dr. Howel Williams and Dr. N. L. Taliaferro at the University of California kindly assisted by loan of equipment for rock-sectioning and for field work. The line drawings and maps were made by Owen J. Poe.

The senior author is responsible for the section on Vertebrate Paleontology, the junior author for the section on Reconnaissance Geology.

Geography and Climate

El Salvador lies southeast of Guatemala and southwest of Honduras, encompassing a roughly rectangular area of 13,176 square miles. Its population is over a million and a half, concentrated in the more arable western two-thirds of the country. A primary system of good roads joins the larger cities, which are located in the central and southern half of the country. Localities off the main routes are accessible by dry-weather road, oxcart, or foot trails.

Rainfall averages about 75 inches a year, but is limited almost entirely to the period between June and November. Since much of the country has been denuded of natural cover, a serious water shortage may develop during the extensive dry season despite the high average rainfall. As is typical in tropical latitudes, temperature has little annual variation, its range being controlled primarily by elevation.

The portion of El Salvador lying east of the lower reaches of the Río Lempa is locally termed "Oriente". This convenient designation is used in this report.

Reconnaissance Geology

General Geology

Topographically, El Salvador may be divided into four general units: (1) A narrow coastal plain; (2) Coastal volcanic belts; (3) A central interior valley area; and (4) A northern series of mountains in part folded and faulted, and in part volcanic, continuous with the Honduran ranges.

Except for a small area of crystallines and Mesozoic sediments in the northwest corner of the republic, all rocks exposed in the localities visited are volcanic or fluviatile deposits of late Tertiary and Quaternary age.

FIGURE 1.—MAP OF EL SALVADOR SHOWING LOCALITIES

FIGURE 2.—MAP OF EL SALVADOR SHOWING TOPOGRAPHIC AND GEOLOGIC FEATURES

Coastal Plain

Between the coastal mountains and the Pacific lies a semicontinuous low coastal plain of variable width. At the extreme western end of the country in Ahuachapan, the plain is narrow where the Santa Ana volcanic group approaches the coast. It broadens to the east, running inland south of Volcán Santa Ana, and then curves back to the coast around the west end of the Balsam Range. The plain is very narrow or absent from south of Ishuatán to La Libertad, whence it broadens gradually to its greatest width of 15 miles at the Río Lempa. East of the Lempa, it narrows somewhat along the south flank of the San Miguel volcanic group, almost vanishes south of the Colinas de Jucuarán, and then again widens around the Golfo de Fonseca.

The coastal plain reflects recent emergence of coastal El Salvador. The magnitude of the emergence is indicated by a marine terrace 10–12 feet high along the beach between Hacienda San Ramon and Intipuca and by two well-preserved river terrace levels, the older 25 feet, and the younger 10 feet above present river level along the Río Goascorán, southeast of Pasaquina. Both localities are in La Union. Consequently, a 25-foot change of level in two stages must have taken place quite recently. Spurs of tuff breccia and basalt which reach the shore along the south flank of the Colinas de Jucuarán near Intipuca have been sharply truncated by wave action at present sea level.

Coastal Volcanic Belts

General description.—The volcanic mountains of the coastal province are divided into an older series consisting of the Balsam Range and the Colinas de Jucuarán, both adjoining the coast, and a younger belt immediately to the north consisting of recent volcanos, some still active, which extends from near the coast in Ahuachapan in the west to the Río San Miguel in the east. The youthful volcanic belt is continuous westward with the young volcanos of southern Guatemala and eastward, by way of young volcanic islands in the Golfo de Fonseca, with Volcan Cosequina and the volcanos of the Marabios Range of southern Nicaragua.

Balsam Range.—This range, so named from the indigenous medicinal balsam trees, borders the coast along the eastern half of western El Salvador. The eastern portion of the range is a maturely dissected volcanic terrain with the original volcanos greatly eroded. No evidence of recent volcanic activity in this area has been noted. The western portion, visited at Hacienda Chilata, is a plateau of pyroclastics, about 500 feet in elevation near the north edge and sloping gradually southward to the sea. Drainage in this portion is by short, south-flowing consequents in deep canyons. Except for a few interbedded lava flows of small volume, the plateau is composed wholly of andesite tuff breccia and interstratified tuffs (Pl. 2, fig. 1). The flows and a majority of breccia fragments are hypersthene andesite. At Hacienda Chilata, the breccia aggregates a minimum of 1,000 feet, thinning to about 400 feet 5 miles southwest at Ishuatan. The plateau surface and underlying stratification dips 3° southwest away from the volcano remnants at Mojon del Izote (a few miles east of Hacienda Chilata), the probable source for the pyroclastics.

The north boundary of the Balsam Range was not examined in detail. However, it rises so abruptly from the Zapotitan basin as to suggest that its north front may be delimited by normal faulting, as is the case in the Colinas de Jucuarán at Lake Olomega. The low area from Lake Zapotitán west to Sonsonate may have developed by subsidence along this boundary.

The destruction of original volcanos by erosion and the relatively low relief of the Balsam Range suggest that it is late Pliocene or early Pleistocene, antedating the youthful volcanic belt to the north. The present deeply incised drainage must be at least in part due to rejuvenation produced by recent coastal emergence.

Colinas de Jucuarán.—This moderately elevated range trends northeast from the mouth of the Río San Miguel to Volcán Conchagua on the Golfo de Fonseca. It is a terrain of volcanic peaks with plateaus and saddles of eruptives joining them eroded to a stage of late youth. As in the Balsam Range, streams of the Colinas show evidence of rejuvenation due to recent uplift. At Lake Olomega, where the north

front of the Colinas was examined in some detail, the volcanic terrain is modified by normal faulting.

The Colinas at Lake Olomega consists of a series of short east–west trending fault ridge with steep north fronts rising steplike southward. The ridges plunge gradually eastward, disappearing in the north-south gap near the village of Tierra Blanca. The lower ridges end abruptly to the west; the higher one is continuous with the irregular volcanic hills in the western Colinas. In addition to topographic evidence, the faulting is indicated by the alignment of a series of hot springs along one fault trend and by dike intrusion along another. The combined vertical displacement of the step-fault series approaches 2,200 feet at Lake Olomega, diminishing rapidly to either side. Lake Olomega probably originated by subsidence along the faults, producing a topographic low which pirated the Río San Miguel eastward. That Lake Olomega was once more extensive is indicated by well-bedded, water-laid sandy tuffs west of Miraflores along the International Railroad of Central America.

Rocks of the Colinas at Lake Olomega and to the south are coarse tuff breccias and interbedded basic flows. These are about 2000 feet thick just South of the lake. The breccia is typically an irregularly bedded jumble of basalt fragments, varying from a quarter of an inch to 4 feet in diameter, in a matrix of buff volcanic sand and lapilli. The common type of interbedded flow is augite hypersthene olivine basalt, the olivine characteristically altering to iddingsite. Identical rocks are exposed along the coast between Hacienda San Ramon and Intipuca; the intervening structure south of the fault series at Lake Olomega dips gently southward and at the coast is nearly horizontal, although with slight undulations.

Immediately east of Tierra Blanca rises an extinct volcano, Cerro Chicarán, which appears to be the source for the major volume of volcanic rocks in the south Olomega area. Its slopes are covered with the same extrusive series that crops out there.

Ten miles east of Lake Olomega, at the intersection of the Colinas de Jucuarán strike with that of the youthful volcanic belt, stands the great extinct Volcán Conchagua, the largest in the Colinas group. It has been eroded extensively and the top is now converted into a gently sloping plain covered with grass and dotted with groves of pines.

The absence of recent activity and the topographic maturity of the Colinas compares closely with the Balsam Range. They probably represent the same period of volcanism in late Pliocene or early Pleistocene.

Youthful volcanic belt.—The most spectacular topographic unit in El Salvador is the late Pleistocene to Recent volcanos. The belt is divisible into several groups, each isolated but all closely related in age and located along a zone of crustal instability paralleling the coast line. This volcanic zone has shown strong seismic activity in recent time; an earthquake totally destroyed San Vicente in 1873, another greatly damaged San Salvador and environs in 1919, and a third damaged San Vicente in 1936.

Volcán Santa Ana Group: This westernmost group forms a massive ridge covering much of Ahuachapan and extending eastward to Lake Coatepeque. It includes the well-preserved volcanic peaks of Apaneca, La Lagunita, San Juan, Las Aguitas, Aguila, Los Naranjos, Santa Ana, Marcellino, and the active Izalco. Izalco, very active in recent years, is reported to have formed completely during the period 1635-1798. In April, 1942, it was erupting dark clouds of dust and cinders at 10–15 minute intervals.

Lake Coatepeque, which lies at the foot of the east slopes of Volcán Santa Ana and Volcán Marcellino, presumably occupies an explosion caldera related to the Santa Ana group.

Volcán San Salvador: This next major member of the recent volcanic range stands isolated a few miles northwest of the capital city. It is separated from the Santa Ana group by the Zapotitan basin. As recently as 1917, an eruption built a small cinder cone in the crater, and lava breached the northwest flank and flowed eight miles down the slope. An elevated peak on the eastern edge of the crater rises several hundred feet above the remaining rim, a remnant of a much larger cone, most of which has been blown away. Volcán San Salvador is a likely source for the pumice tuffs which surround Lake Ilopango and underlie the capital city.

Lake Ilopango: This lake, located half-way between Volcán San Salvador and Volcán San

Vicente, is volcanic in origin. The mechanism of its formation is uncertain, but emergence of a central volcanic island in 1880 shows it to be still a locus of activity. The location and recency of activity are consistent with the youthful volcanic belt, and the lake appears to represent the caldera of a member that was completely destroyed by explosion. Remnants of a cone, if present, are now buried by later pumice tuff.

VOLCÁN SAN VICENTE: Southwest of the town of San Vicente is the great double-peaked Volcán San Vicente. The two cones are still well preserved. No recent activity other than that of fumaroles has been recorded, but the youthful topography indicates it to be but recently extinct or perhaps only quiescent. The volcanic belt is broken east of San Vicente by the Rio Lempa valley.

VOLCÁN SAN MIGUEL GROUP: The San Miquel group forms a continuous range between the Río Lempa and the Río San Miguel. The cones, joined by high saddles of eruptives, are all Recent to Subrecent, as judged by their preservation. Volcán San Miquel is a perfect cone with a cloud of sulphur dioxide and steam hovering almost continuously above the crater. It has erupted repeatedly in recent time, most intensely in 1787 and 1844. A small fresh basalt flow on the northwest flank, still devoid of vegetation, dates from the 1844 eruption.

Central Interior Valley Area

Definition.—A broad, relatively low area between the Coastal Volcanic Range and the Northern Ranges extends the length of the country. This central area is elevated to the west and slopes eastward to near sea level at the Golfo de Fonseca.

Volcanic plateaus.—Associated with and derived from the youthful volcanic belt are extensive plateaus of eruptive rocks, predominantly pumice tuff and breccias with an interbedding of flow rock along the volcano slopes. The major tuff plateaus lie west of the Río Lempa, one along the north side of the Volcán Santa Ana group from the Guatemalan border to Chilamatal, and a second from Volcán San Salvador eastward to Volcán San Vicente (Pl. 2, fig. 2) and north toward Ilobasco. The tuffs are characteristically light weight, gray to white, massive deposits filled with scattered fragments of fresh white pumice. Particularly good exposures are found in San Salvador and in the area around Lake Ilopango.

The tuffs appear to be the peripheral deposits produced by the youthful volcanic belt and are probably late Pleistocene. The remains of a mastodon were said to have been removed from the tuffs exposed in a road cut near Chilamatal in the construction of that part of the Pan-American highway, but the bones had been badly shattered by a dynamite charge and were never accurately identified. An age older than late Pleistocene is improbable, since the tuffs are quite fresh and form geomorphically youthful plateaus; drainage has carved deep and steep-walled canyons in the soft materials. The tuffs have been undisturbed for the most part since deposition and lie as a horizontal blanket over older volcanic rocks.

Outcroppings of lacustrine or estuarine water-laid tuffs and tuffaceous sand northeast of Volcán San Vicente in Barranca El Sisimico have yielded abundant but poorly preserved leaf impressions and small fishes. The beds at El Sisimico are about 100 feet thick and are overlain by 300 feet of sandy tuff breccia topped by flow rocks. This series, which is gently folded at Barranca El Sisimico, may be the correlative of the tuff plateau forming eruptives. The folding is related to recent faulting. A sharp scarp along the north side of San Vicente Valley, clearly visible from the air, breaks the tuff plateau and indicates very recent faulting along which the San Vicente Valley has been dropped as a graben. Such subsidence by normal faulting must represent a readjustment following withdrawal of vast quantities of magma from beneath the area adjacent to the active coastal volcanos.

Lake Guija Volcanics.—Interior El Salvador from just north of Santa Ana to Metapán and from the Río Lempa to the Guatemalan border is an area of young volcanos and lava plateaus equivalent in age to the youthful volcanic belt. The volcanos, except Volcán Chingo, which reaches an elevation of 5,648 feet, are not as large as those in the coastal belt, nor do they show any obvious alignment. Lake Guija is bounded by low extinct volcanos on the south and east, and its origin is probably attributable to a volcanic damming of the drainage. East of Lake Guija, volcanos are scattered over the

terrain as far as the Río Lempa, and Volcán Masahuat stands within and controls the great westerly loop of the river. An extensive young cinder cone basin occurs to the west of Lake Guija; it covers the El Salvador border region and extends about 20 miles into Guatemala.

Basaltic lavas comprise the bulk of the Guija volcanics with additional minor volumes of basaltic pyroclastics. The profiles of most of the Guija volcanos, lower and more rounded than those of the youthful volcanic belt, reflect this excess of flow material over pyroclastic and perhaps also derivation from a magma more basic on the average. Extremely fresh ropy scoriaceous lavas observed on Hacienda Ostua and along the Río Tahuilapa have been erupted very recently. The Guija volcanics lap up onto intrusives and Jurassic and Cretaceous sediments to the north of Metapán along the base of the high folded mountains along the El Salvador, Guatemala and Honduras border.

Guazapa Area.—Central El Salvador along the longitudinal portion of the Rio Lempa, visited but briefly in crossing from San Salvador to La Palma, is a flat flood plain from immediately north of Guazapa to a few miles beyond the Río Lempa. Its level expanse is broken only by the extinct Volcán Guazapa and related smaller cones. The east-west extent of the flood plain has not been mapped.

Central Oriente.—East of the Río Lempa in the central interior, except for a continuous east-west line of hills roughly down the center of the low, is an area of rolling hills with 200–300 feet relief. South of Jocoro, the central hills reach a maximum elevation of about 1,000 feet above drainage level.

The surface rocks in the vicinity of Divisidero and Jocoro are a series of andesite and basalt flows, basalt tuffs, and tuff breccias, but because the series is poorly bedded, the structure was not worked out. There is evidence of two volcanic series in the area—(1) the underlying folded Pliocene flows, and (2) a locally distributed, overlying series of coarse tuff breccias and interbedded basalt flows in general flat to gently dipping with structure appear dependent on normal faulting. Several rhyolite plugs presumed to be eroded volcanic necks are associated with the younger series.

Pliocene Series: This series, underlying the central Oriente between Divisidero and Gotera, appears to be older than the Balsam Range and Colinas de Jucuarán volcanics. It embraces an undetermined thickness of adesitic flows, agglomerates, tuff breccias, and minor pure tuffs, with the flow rocks predominating. Samples from the Montecristo mines near Divisidero and from along the Gotera road are augite andesites considerably altered through calcitization and epidotization. The Montecristo mines work quartz veins, carrying gold and silver values, which have been emplaced along a system of east-west faults with 40–60 degrees north hade.

Younger Series: Above the Pliocene volcanics are 150–400 feet of coarse basaltic tuff breccia overlain by as much as 500 feet of augite olivine basalt. The breccias are massive to poorly bedded with coarse, gray hypersthene andesite fragments scattered through a matrix of basaltic tuff. About 1½ miles southwest of Divisidero, two small tridymite-rich biotite rhyolite volcanic necks protrude through the breccia. The younger volcanics crop out as the higher hill belt with an abrupt north-facing scarp from Divisidero eastward to near Pasaquina. Evidence of faulting is present along the base of the scarp south of Jocoro, and the present distribution of the younger series is probably delineated by east-west normal faulting. The younger breccia and flow rocks appear to have been derived locally; the volcanic necks represent the probable loci of volcanic activity. These rocks rest on an irregular surface of folded Pliocene volcanics and have since been slightly tilted by normal faulting. The rock types and structure suggest correlation with the Volcán Cacaguatique volcanics to the north of Gotera and with those in the Balsam Range and Colinas de Jucuarán, all thought to represent late Pliocene or early Pleistocene volcanism.

The late Pleistocene mammalian assemblage found in an old boghole resting on weathered andesite at the Hormiguero locality is more recent than the younger series just discussed. The fauna is considered equivalent in age to part of the latest volcanics of the coastal volcanic belts and of the Lake Guija area.

Northern Ranges

Volcán Cacaguatique.—At the northern limit of the Central Oriente lowland is Volcán Caca-

guatique, an extinct volcano 5300 feet high. It covers most of the area south of the Río Torola between the Río Lempa and the headwaters of the Río San Miguel. Structurally, Volcán Cacaguatique is an eroded cone, the original crater having been greatly enlarged by interior drainage which breached the wall to the southeast. The north rim is the highest, probably because it lies in the humid cloud forest zone facing the moisture-laden north winds and consequently has a heavy protective cover of vegetation.

The last eruptions from Cacaguatique are represented by about 1400 feet of volcanic rocks now exposed in the north wall of the old crater. The sequence is composed of a series of basalt flows aggregating 800 feet, which are underlain by 500–600 feet of bedded tuff breccia with interbedded thin flows and tuff. The basalts are fresh, dense, highly jointed porphyries with phenocrysts averaging about 1 millimeter long. Microscopically they show a pilotaxitic to hyalopilitic groundmass with abundant labradorite, augite, and hypersthene phenocrysts. Augite is found in a few crystals as borders around hypersthene kernels. The basaltic tuff breccia looks identical to the coarse breccias in the younger series described from the Divisidero and the Lake Olomega areas.

These volcanics show 15–20 degrees initial dip around the old crater wall, flattening peripherally to 5–10 degrees dip. The tuff breccias are more extensive than the upper flows, and northeast of Ciudad Barrios they crop out as a shelf at least a mile wide. Below the tuff breccias on the northwest slope of Mount Cacaguatique between Ciudad Barrios and the Río Torola, is a bsaic flow series, aggregating 1,500–2,000 feet in thickness, which may be equivalent in age to the Pliocene Series previously described.

Torola Valley.—North of Volcán Cacaguatique lies the east-west trending Torola Valley, the north side of which is formed by folded mountains continuous with the Honduran ranges. The valley is structurally a syncline, the axial portion being occupied by the subsequent Río Torola. The river enters El Salvador about 20 miles northeast of Volcán Cacaguatique and, swinging sharply, flows westward along the syncline axis. In the Torola Valley east of Cacaguatique, erosion has stripped away softer volcanics and left broad areas as scarp-bounded dip surfaces on the valley floor. The syncline limbs dip gently and regularly at 5–10 degrees except where the north limb is broken by a minor anticline which carries through near Jocoaitique. Folded in the syncline is a series of basalt and andesite flows, tuffs and tuff breccias probably largely derived from eruptions in Honduras. South of Meanguera village, a 100-foot unit of buff cross-bedded, medium to coarse arkosic sandstone represents river deposits interbedded with the volcanics. The entire series is thought to be equivalent to the "... nearly 3,000 feet of andesites, tuffs, agglomerates, and interbedded deposits of sandstones, clays, and poorly consolidated conglomerate [of southwestern Honduras] ..." mentioned by Weaver in Schuchert's Historical Geology of the Antillean-Caribbean Region (1935), and to the Pliocene Series of this paper.

Fifteen miles down the Torola Valley, immediately west of Carolina, are exposed thin, fresh-water (probably lacustrine) limestone, chert, and hard, gray, cross-bedded, medium to coarse, pebbly sandstone with a few bone fragments. These may be in general equivalent to the sandstone deposition at Meanguera. A portion of a femur, a few other bone fragments, and a poorly preserved mold of a mastodon tooth were found in the sandstone in Río Chula, but unfortunately were so incomplete that generic determination was impossible. The remains might indicate either Pliocene or Pleistocene age.

Metapán Area.—North of the Lake Guija volcanic area towers the highest range of El Salvador, reaching a hieght of 9,200 feet at Los Esesmiles. Intrusive monzonite, Mesozoic sediments, and Pliocene volcanics are exposed north of Metapán in the extreme northwest corner of the country. An east-west line through Metapán roughly demarcates the older rocks to the north and the young volcanic cover to the south. Exposures of the older rocks were observed also at Hacienda San Miguel and at San Juan, about 10 miles east of Metapán.

The section at Hacienda San Miguel is as follows:

Age	*Formation and lithologic character*	*Approximate thickness (feet)*
?Pliocene	Buff, coarse, poorly consolidated sandstone and cobble conglomerate, average diameter of cobbles is 4 inches. Phenoclast 70% andesitic porphyry, 30% reworked middle Albian limestone.	650
	Unconformity	
?Pliocene	Gray andesite	30
	Unconformity	
Cretaceous Middle Albian	Massive, cave-forming, dense, blue-gray, poorly fossiliferous limestone. One *Exogyra* sp.	400
	Metapan beds:	
Lower and Middle Jurassic	1. Red brown shale.	180
	2. Hard vein quartz pebble conglomerate.	30
	3. Interbedded fine-grained green (epidote rich) sandstone and red, brown and green shales.	75+

The Metapán beds, which outcrop widely in the Sierra de Metapán of El Salvador and in adjoining Guatemala, are lithologically similar to the Todos Santos beds of northern Guatemala and Chiapas, and have been correlated with them by Sapper (1937), who considers them to be probably equivalent also to the upper part of the Tegucigalpa formation of Honduras (also see Imlay, 1944). Mullerried (1939) found Lower and Middle Jurassic plants in the Metapán beds. He has noted occurrences of Lower Cretaceous Neocomian shales, marls, and limestones in the upper part of the Todos Santos beds and suggests that upper Metapán and Tegucigalpa beds may be of the same age. The Metapán beds, as exposed at Hacienda San Miguel, are the highly colored nonmarine types which Mullerried cites as typical of the Lower and Middle Jurassic. Beds which might be equivalent to the calcareous Neocomian part of the Todos Santos beds are absent. Instead, Middle Albian limestone directly overlies with apparent conformity the nonmarine Metapán beds, which indicates a long hiatus between the Metapán beds and the Middle Albian limestone in this area and suggests an unconformity through overlap.

Mullerried (1939) collected *Toucasia* cf. *texana* Roemer from the massive limestone 1 kilometer north of Metapán, dating it as Middle Albian. In addition to the above occurrences at Hacienda San Miguel and north of Metapán, the limestone crops out extensively northwest and west of Metapán. In the latter areas, scattered quartz veins cutting it carry minor amount of lead and copper and one vein has been worked for small gold and silver values.

At Hacienda San Miguel a thin andesite flow followed by sandstone and heavy conglomerate unconformably overlies the limestone. The conglomerate cobbles are derived from both the andesite and limestone. To the east and northeast thick andesite tuff and flow lie stratigraphically above the conglomerate. No fossils have been found in the conglomerate or tuff. However, they are thought to represent the easterly extension of the volcanic series with interbedded continental sediment described by Weaver in southwest Honduras, and are referred to Pliocene.

At the San Juan Mine, a small galena and sphalerite replacement zone in the Middle Albian limestone has been prospected. Here the limestone is in probable fault contact with an extensive monzonite body, which forms the flooring rock of the lower slopes of the Sierra de Metapán and the fronting flat south to the Río Tahuilapa. The subrecent eruptives of the Guija volcanic basin cover all older rocks south of the Río Tahuilapa. The monzonite at San Juan is a buff, coarsely crystalline rock as exposed at the lower elevations, but finer-textured with more elongate crystals higher up the slope where cooling was rapid on the periphery of the intrusion. Mineral composition, as determined from several thin sections, is 40–50% orthoclase, 25–30% plagioclase (ca. oligoclase), 15–20% quartz, 10% pleochroic hornblende, minor biotite and magnetite. Many of the ferromagnesians have altered to calcite and chlorite. A plug-like ore body about 50 yards in diameter in the Tejada mine appears to have been derived as a segregation from the monzonite. Specular hematite and secondary limonite form the bulk of the deposit, with minor hemimorphite and a showing of copper as

secondary carbonates. Early Spaniards worked the Tejada mine to obtain iron for tools.

Intrusive monzonite identical in composition to that at San Juan crops out on the northeast slope of the Metapán range and was also seen in a tiny fault sliver in the Río Gramal east of La Palma. It is doubtless exposed more widely in the southeast portino of the Los Esesmiles mass, but was not encountered in the western portion above San Ignacio and La Palma. These scattered occurrences indicate that monzonite may underlie a large part of the northern Occidente at relatively shallow depth. No definite age has been established for the granite, but we assume that it is pre-Mesozoic.

Overlying the monzonite on the slope northeast of San Juan is a massive, highly jointed, light-colored andesite tuff with a minimum thickness of 400 feet. Thin sections of this vitro-crystalline rock reveal scattered kaolinized orthoclase, quartz, spongy hornblende, and rare plagioclase crystals. The tuff continues eastward for several miles, forming steep cliffs above the monzonite. A similar rock is exposed on the northeast side of the Metapán range halfways between Hacienda San Miguel and Citalá. There also it rests on monzonite and is overlain by 500 feet of massive andesite, which forms the higher prominences of the range crest in its eastern portion.

At the base of the steep south slope of the Sierra de Metapán, north and east of Metapán, 200–300 feet of poorly consolidated conglomerates, gravels, and sands have been deposited as an alluvial apron in Pleistocene and recent time.

Los Esesmiles and the Sierra Madre.—East of the Río Lempa, the Sierra de Metapán structure continues in the Sierra Madre and Los Esesmiles areas. Los Esesmiles was visited briefly northeast of San Ignacio, where the exposed section is entirely volcanic, aggregating as much as 3,000 feet of intermediate to basic flows, tuffs, and tuff breccias. The Sierra Madre range, as crossed on the highway from La Palma to Tejutla, is also comprised of volcanics similar to and probably of the same series as that at Los Esesmiles. No direct evidence as to their age has been found. They presumably are continuous eastward with the Pliocene volcanics of the central and northern Oriente; freshness of much of the rock is consistent with such late age.

The south and east portions of Chalatanango and the Department of Cabañas were not visisted. However, the Sierra Madre Range is known to extend southeasterly across Chalatanango as a strongly elevated bulwark to the junction of the Ríos Lempa and Sumpul. In Cabañas, the range rises to a lesser elevation similar to the highland trending northwest from Volcán Cacaguatique. Over this entire distance, the bedrock is reported to be volcanic, and it is probable that the Pliocene volcanic series is exposed continuously from the northern Oriente to the Sierra de Metapán.

Geologic Summary

The oldest rocks in El Salvador are those of the Sierra de Metapán. The monzonite intrusion exposed at San Juan is undated, but is thought to be pre-Mesozoic. In probable fault contact with the monzonite is a section of nonmarine Metapán beds, dated as Lower and Middle Jurassic by Mullerried, overlain by limestone from which Mullerried collected a Middle Albian fossil. An unconformity between the two is indicated.

The Mesozoic rocks are unconformably overlain by a thin andesite followed by sandstone and heavy conglomerate; the conglomerate has been in part derived from the Cretaceous limestone. Stratigraphically above the conglomerate as exposed to the northeast and east is andesite tuff and flow. Direct evidence as to the age of the conglomerate and andesite series is lacking. They are a part of a thick sequence of andesite and basaltic flows, tuffs, and breccias, with minor interbedded continental sediments, which form the Northern Range of Los Esesmiles and western Chalatanango and apparently continue eastward across Chalatanango and Cabañas into the northern Oriente. These are considered to be the same volcanic series as Weaver (Schuchert, 1935) described in southwestern Honduras and are questionably Pliocene. The Pliocene series which underlies the central Oriente from Divisidero northward, and the thick volcanics with minor nonmarine sediment north and south of the Río Torola, upon which Volcán Cacaguatique is superimposed, are likewise considered to belong to this great

series. Indeterminable mastodon remains at Río Chula suggest Pliocene or early Pleistocene age. Folding, faulting, and mineralization have effected these rocks in the Oriente, and the mineralization near Metapán may be of equivalent age.

Eruptives of the Balsam Range, Colinas de Jucuarán, Volcán Cacaguatique, and the younger series of the central Oriente are all grouped as probable late Pliocene and early Pleistocene. The partial to almost complete erosional obliteration of the source volcanos in these groups point of such an age. Hypersthene andesite, hypersthene and olivine-bearing basalt flows, and a vast amount of coarse tuff breccia are the characteristic rocks. Along the north front of the Colinas de Jucuarán, these rocks have suffered normal faulting.

The youngest rocks in El Salvador are eruptives from the youthful volcanic belt and from the volcanos and cinder cones of Lake Guija. These are largely late Pleistocene, but in the youthful belt minor volcanic activity still continues. The eruptives are basaltic and lesser andesitic flows, tuffs, and lapilli. The extensive pumice tuff plateaus along the north flank of the Volcán Santa Ana group and at San Salvador and vicinity were derived from the youthful volcanic belt, probably during late Pleistocene. The late Pleistocene fauna from a surficial bog deposit at Hormiguero is considered correlative to a part of these youngest volcanics.

VERTEBRATE PALEONTOLOGY

Pleistocene Faunal List

Mammalia
- Edentata
  - Megatheriidae—*Megatherium* (giant ground sloth)
- Carnivora
  - Felidae—*Smilodon* (sabre-tooth cat)
- Notoungulata
  - Toxodontidae—Gen. et sp. indet. (toxodont)
- Proboscidea
  - Gomphotheriidae—*Cuvieronius* (spiral-tusked mastodon)
  - Elephantidae—*Mammuthus* cf. *jeffersonii* (mammoth)
- Artiodactyla
  - Camelidae—small camelid
  - Bovidae—*Bison* (bison)

?Pliocene or ?Early Pleistocene

Río Chula.—Our attention was directed to some mastodon remains which Señor Fredrico Portillo, of San Miguel, had found near the village of Carolina, Departemento de San Miguel (Fig. 1). The locality occurred in the stream bed of the Río Chula near the small village of Tovos (El Caserio de los Tovos) on the property of Margarita Vde Portillo, in Canton Nacaspilo. The fossils were found in a bed of well-indurated sandstone where a peon ruined them in trying to gouge them out. The sandstones and limestone cherts in which the fossils occur were interbedded with the lava flows of the area. The fossil member was not extensive enough to work for bones, nor were the fragments left in the rock worth collecting.

FIGURE 3.—MAPS SHOWING LOCATION OF HORMIGUERO FOSSIL QUARRY

Señor Portillo told us that a well preserved mastodon tooth had been taken from this locality and sent to the National Museum in San Salvador, but we were unable to find any

record of it at the museum. From Portillo's description of the specimen, we have concluded that it is not older than Pliocene and could be early Pleistocene.

possible to collect the fossils at that time, the site was not visited.

According to Mr. Swanquist, a local farmer snagged a large bone with his wooden plow

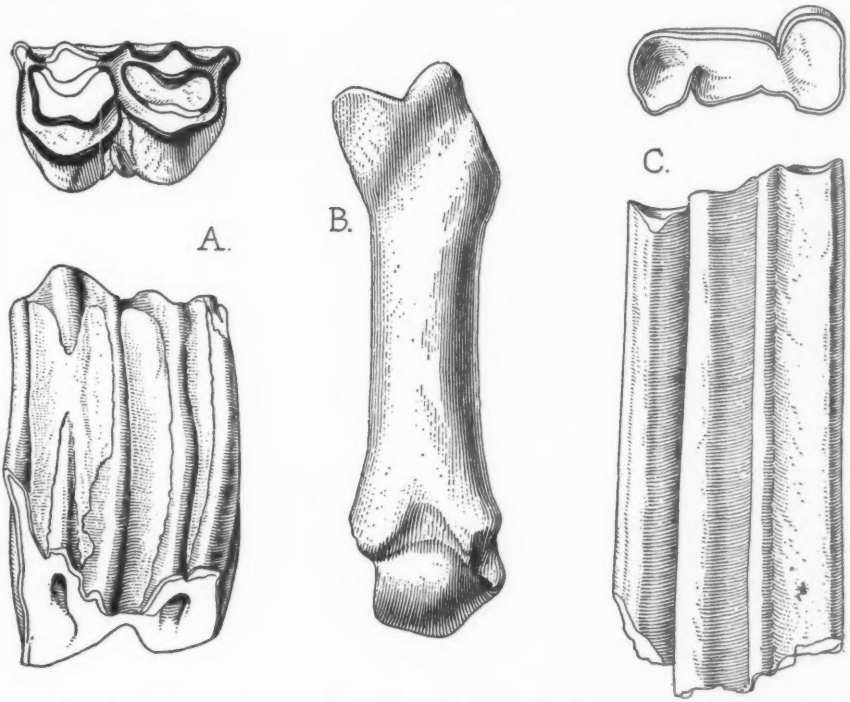

FIGURE 4. BISON UPPER MOLAR, SMILODON RIGHT METACARPAL II, TOXODONT LOWER MOLAR
(A) *Bison*, upper molar, occlusal and labial views, UC 37035; (B) *Smilodon*, right metacarpal II, front view, UC 37036; (C) *Toxodontidae*, gen. et sp. indet. occlusal and lingual views, UC 37040. Hormiguero fossils × 1, late Pleistocene, loc. V4201, El Salvador, Central America.

Late Pleistocene

Hormiguero.—A large concentration of giant ground sloth bones was uncovered at the Hormiguero fossil vertebrate locality on the Hacienda San Juan del Sur in the Department of San Miguel. The hacienda is the property of Señor Francisco Fuentes and, at the time our collecting was being done, was under the management of his brother, Señor Florentin Fuentes.

The senior author's attention was first directed to this deposit by Mr. G. A. Swanquist in 1925, when a bird and mammal survey of El Salvador was being conducted. Since it was not

while preparing the ground in this area for the production of his annual corn crop. Some local person became interested in the remains of the so-called giant, opened a quarry and removed a large number of bones. These were taken via carreta to the town of Gotera, there they were heaped in the village square where all interested persons could see and possibly carry home a piece of broken bone as a souvenir. There is no trace of that original collection today.

Our first visit to the locality was on December 21, 1941. After some inquiry at the headquarters of the hacienda, a man came forward

and told us that he as a boy remembered seeing them dig up bones of the giant. He then led us to the spot near a sticky mudhole that had nearly dried up (this being near the end of the dry season). Though the site of the old quarry was overgrown with brush, the outline of the excavation was clear (Pl. 1, fig. 1). The exact location is 50 yards south of the Morozan-San Miguel department line and 350 yards west of the Hormiguero-Divisadero camino on the south side of a small ravine (fig. 3).

We returned to the site on April 13, 1942, and worked the quarry 10 days. With our collecting materials used up and the rainy season upon us, we closed the quarry; but in this time we had collected and prepared for shipment five large collecting boxes of fossil remains, mostly of the giant ground sloth *Megatherium*. Remains of other mammals were rare. These included an upper molar of *Bison*, a lower molar of a toxodont, the right metacarpal II of *Smilodon*, a proboscidean patella and fibula, and a small camelid vertebra that was too poorly preserved to collect.

The fossils were found 5–6 feet below the surface in a layer of sticky gray mud. The andesite bed rock showed an old weathered surface usually about 3 inches thick. This clay varied from red and yellow to light gray. Above was the bone level of gray andesitic clay, varying from 12–18 inches in thickness. Mixed with the bones, and in some instances in the weathered andesitic clay below, were rather hard andesitic stones ranging from 1–6 inches in diameter. The vertical section above the bone level in the quarry showed a slightly banded or laminated effect (Pl. 1, fig. 3, 4)—9 inches of black clay, 6 inches of dark-gray clay, 9 inches of blue-gray clay, and 21 inches of gray clay, which graded up into the top soil. In the base of the top layer were some rounded andesitic stones less numerous than those in the bone layer. No bones were found in the three upper layers. There was lateral variation in the thickness of the different layers, some layers may have lensed out beyond the edges of the quarry.

An occasional bone was discovered in the black clay overlying the bone layer, but bones were so thick in the lower level that it was necessary to sacrifice most of the incomplete or poorly preserved bones in the process of blocking out the better specimens. Crania and mandibles were badly broken; all that we found were fragments and usually the teeth had fallen out of the alveoli. Only two or three ribs were unbroken. Other more delicate bones, especially those of the younger animals, were fractured or unbroken.

The region about Hormiguero, when *Megatherium* and its contemporaries were common, was probably covered with a deciduous forest much heavier in the lowlands and valleys than on the ridges, where the soil mantle was thinner. This forest was possibly dominated by *Ceiba*, *Cecropia*, and other giant trees, which may be seen in undisturbed tracts of virgin forest near Lake Olomega and in the Colinas de Jucuarán. We assume that large cumbersome brutes like *Megatherium* favored the more heavily forested areas in the valleys and lowlands where the terrain was not so favorable for the growth of tangled and thorny undergrowth, due to the dense canopy of foliage 60 or more feet above the forest floor. Dickey and van Rossem (1938, p. 31–33) give a clear picture of the climatic conditions that prevail in this region today. Their Arid Lower Tropical life zone extends from sea level to an elevation of 2500 feet. The seasons in this life zone are marked by 6 months of rain and 6 months of drouth. The dry season evidently was a critical period in the accumulation and preservation of the Hormiguero fossil remains. At that time (March and April in El Salvador) only the larger streams and a few springs in the hills provided water for the larger terrestrial animals.

Faulting or jointing in surface rocks probably concentrated underground water until it seeped or flowed out through the andesite at the site of the Hormiguero locality. As indicated by similar springs in the vicinity today, this was probably a well-supplied source of water that prevailed throughout the dry season and, therefore, became a waterhole at the edge of a stream bed. The main drainage then shifted to the east. The quarry site became isolated and filled slowly with an accumulation of humus and very fine decomposed andesitic silt, thus developing into a boghole. Since the bones display very little or no abrasion, and in most instances fairly sharp breaks are still well-

FIGURE 1. AFTER THE BRUSH HAD BEEN CLEARED AWAY

FIGURE 2. DISTAL PHALANX OF MEGATHERIUM

FIGURE 3. MEGATHERIUM LIMB BONES AS THEY WERE BEING UNCOVERED BELOW THE LAMINATED SEDIMENTS

FIGURE 4. COLOR LAMINATION ALONG THE QUARRY WALL AND THE ACCUMULATION OF BONES

HORMIGUERO QUARRY SITE, EL SALVADOR, CENTRAL AMERICA

FIGURE 1. TYPICAL TUFF BRECCIA OF THE WESTERN BALSAM RANGE
Rio Ayacachapa near Hacienda Chilata.

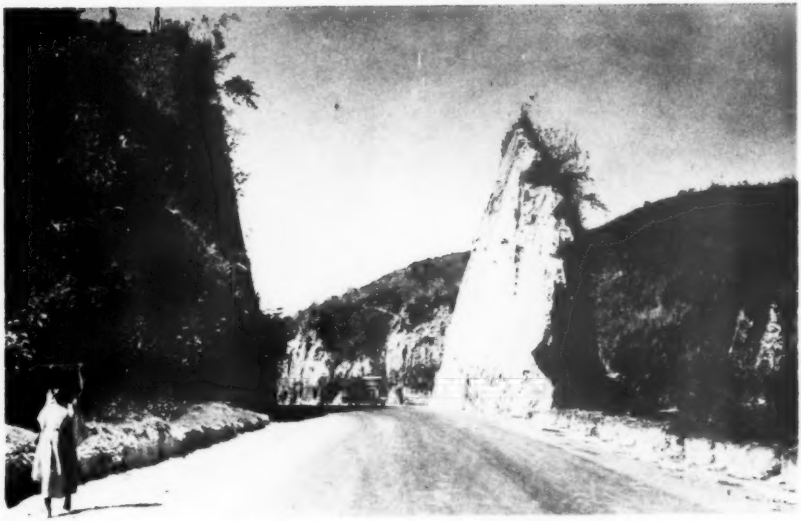

FIGURE 2. ZACATECOLUCA-SAN SALVADOR HIGHWAY
Showing a typical tuffaceous exposure in youthful volcanic belt.

PLEISTOCENE TUFFACEOUS EXPOSURES, EL SALVADOR, CENTRAL AMERICA

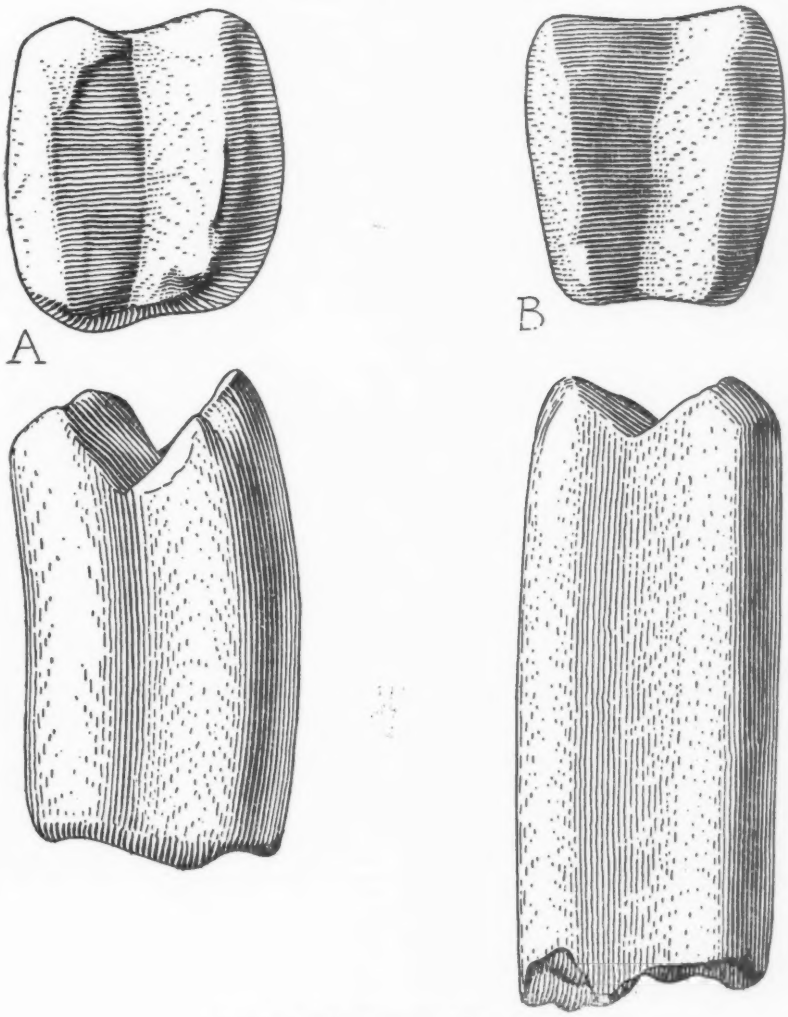

FIGURE 5.—MEGATHERIUM CHEEK TEETH

With lower parts broken off, occlusal and lateral views, UC 36996 and UC 37000. Hormiguero fossils × 1, late Pleistocene, loc. V4201, El Salvador, Central America.

preserved, and because the different elements are completely disassociated yet thickly matted in the bottom of the deposit, it seems that the animals became trapped in the mud while attempting to get a drink. There they struggled, all the while becoming more securely entangled in the sticky mud. An animal caught in this manner would die in a few days, leaving the carcass partly buried and partly exposed. In the succeeding dry season, or even later in the same season, these bones were trampled by other animals caught in the same manner. Thus the bones became disassociated as they sank or were pushed deeper into the mud.

Megatherium with a heavy body and short limbs, and with a hind foot over 2 feet in length, would be more susceptible than the average animal to these death traps. The old and the young alike were caught and buried. Other animals like the bison, the toxodont, the camelid, and the sabre-tooth cat probably tended to occupy the more open but limited grassland areas where they depended on the larger streams for their water supply. The few that occasionally frequented these bogholes were lighter or at least less clumsy and hence less liable to entrapment than the ground sloths.

If these bones had been washed and deposited into this pocket by the action of a stream, a current of such carrying power also would have transported considerable sand and gravel, of which there is no trace. The few cobbles in the basal layer with the bones are possibly the residuals in a relatively clean-swept bend of the old stream bed where they came to rest before the stream channel shifted.

We can offer no satisfactory explanation for the color lam nation in the quarry (Pl. 1, fig. 4). The texture of the sediments appears the same in the different layers, and there is only a slight difference in the color. It does not appear to be a stratigraphic phenomenon.

This means of entrapment and accumulation of bones of Pleistocene mammals, especially *Megatherium*, is frequently encountered in the tropical regions of America.

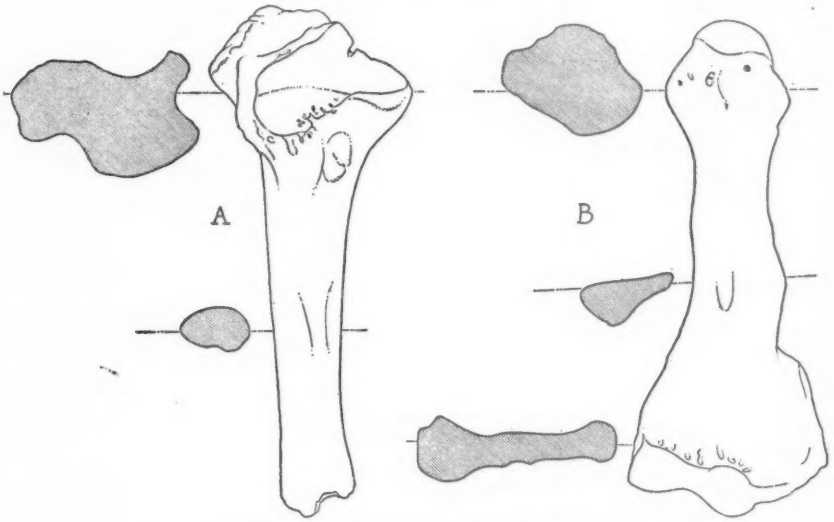

FIGURE 6.—MEGATHERIUM ULNA AND HUMERUS

(A) ulna, front view and cross sections, distal end broken off, UC 36888; (B) humerus, front view and cross sections, UC 36886. Hormiguero fossils × $\frac{1}{10}$ late Pleistocene, loc. V4201, El Salvador, Central America.

"There is in the Chicago Natural History Museum a collection from a comparable, Central American bog deposit. This was found by Paul O. McGrew and Albert Potter in the Province of Copan, Honduras. Conditions of deposition are generally, although not exactly, comparable, preservation is identical and faunal composition closely comparable to Hormiguero. The faunal list is as follows: *Megatherium*, glyptodont, *Felis concolor*, toxodontid, proboscidean, *Equus*, *Camelops*?, *Odocoileus*?.

"*Felis*, *Camelops*? and *Odocoileus*? are represented by one bone of each. It is impossible to say how many individuals of the others there were. Megathere remains greatly outnumbered all the rest, there being at least three adults and one juvenile present. At least two horses are represented. The proboscidean bones are all immature and very probably from one individual. The age must be almost identical to that of Hormiguero, a point of interest in connection with the notes on the megatheres that follow." (Bryan Patterson, personal written communication.)

We have concluded that the age of the mammals in the Hormiguero quarry is late Pleis-

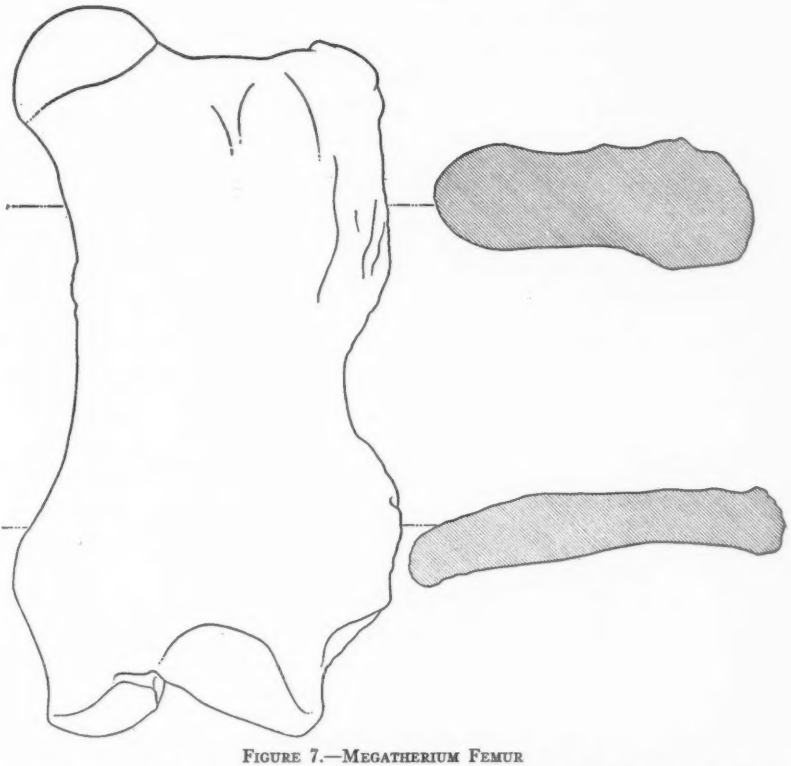

FIGURE 7.—MEGATHERIUM FEMUR

Front view and cross sections, UC 36884. Hormiguero fossils × $\frac{1}{8}$, late Pleistocene, loc. V4201, El Salvador, Central America.

tocene. The deposit in which the bone occurs is horizontal in position and the late date of the bed is further indicated by the fact that part of the boghole is still active. There was evidence of a horse having been caught there a few months before our arrival. The presence of *Bison*, usually considered as middle to late Pleistocene arrivals from the Old World, also tends to support this point of view though, obviously, not definitive of even Wisconsin time. The bones, too, though in an excellent position to be replaced, are not as heavy as fresh bone. It is quite possible that these animals existed in El Salvador in Wisconsin time and we should not be surprised if future discoveries reveal Sub-Recent stragglers.

Other Pleistocene Localities.—Large Pleistocene mammals evidently were common in El Salvador as indicated by a list of localities supplied us by Dr. Dagoberto G. Gavisia, the El Salvador Consul General in San Francisco, and by Dr. Mario Lewy van Severen, Chief in the Department of Chemical Agriculture and Animal Nutrition. They found in their National Museum records 20 localities from 8 of the 14 departments in El Salvador. Five of these finds were made in the Departemento de Usulutan.

We did not find more fossil bones at any of these localities and could not check the authenticity of these records; nor could we locate the original sites. Evidently, though, with one or two exceptions, these discoveries represent the isolated remains of large mammals that

became buried in sites suitable for their eventual preservation as fossils. Most of the specimens are now lost, but a few remain in the National Museum in San Salvador though their exact localities are not recorded.

vador. No one has a clear conception of individual variation in these large animals. Two characters in our specimens may or may not be significant, one is the large and flattened ungual phalanx "of the third digit of the manus" (Bryan Patterson) like that in *M. mirable* Leidy and in *M. hudsoni* White;

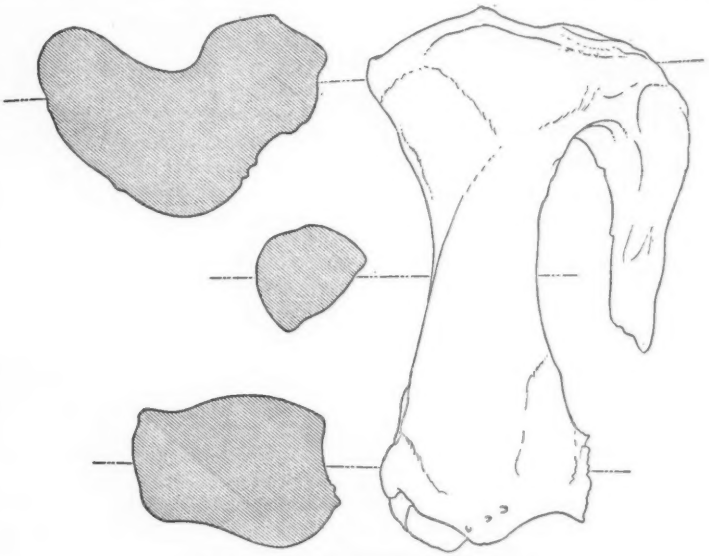

FIGURE 8.—MEGATHERIUM TIBIA-FIBULA

Front view and cross sections, distal end of fibula broken off, UC 36885. Hormiguero fossils × $\frac{1}{8}$, late Pleistocene, loc. V4201, El Salvador, Central America.

Description and Discussion of Fossil Vertebrates

EDENTATA

MEGATHERIIDAE

Remains of the giant ground sloth *Megatherium* (Figs. 5–10) are more frequently encountered than are the bones of other Mammalia in El Salvador. Many of the finds commonly referred to as "Mastodonte" are usually referable to *Megatherium*. Bones of these huge creatures were matted in the bottom of the Hormiguero fossil quarry (Pl. 1, figs. 3–4). Unfortunately only fragments of crania and mandibles were found, and though we have nearly every part of the body skeleton represented in our collection, specific identification of the materials is impossible at this time.

Eleven species or more of *Megatherium* have been described from North and South America, but, for the most part, the diagnoses are inadequate for the identification of materials like those from El Salvador and the other is the relatively small humerus as compared with the large femur. We have observed by a detailed comparison with the figures and descriptions of Owen (1855; 1856; 1858; 1859) that the El Salvador bones are remarkably close to those of *M. americanum* Cuvier except for the large distal phalanx which is flattened and measures $13\frac{1}{16}$ inches (331.5 mm.) in length. Though the bony sheath is missing as indicated by Patterson (see comments below), our specimen is not crushed nor is it an artifact (Pl. 1, fig. 2; Fig. 9A).

The Megatheriidae are of South American origin and first appear as distinct from the other ground sloths in the Miocene of Argentina. One of the first representatives to reach North America *according to White* (1941) is *M. hudsoni* from the Bone Valley Middle Pliocene of Florida. No other megatheres have been recorded from the Pliocene in North America, and the meager materials from Florida are surprisingly like the Pleistocene remains from that state and elsewhere. The other recorded ground sloth specimens of undoubted Middle Pliocene age

Comparative measurements in humeri and femora in Owen's type, in the Hormiguero and in the Honduras specimens

	Humeri					Femora			
	CNHM			Owen	UC36886**	CNHM		Owen	UC36884**
	P27080	P27081	P26970			P26970†	P27080		
Length	751	734	795	760*	763	687	706	690*	898
Proximal width	208	197	210	—	192	346	354	—	456
Width at center of shaft	146	150	155	—	142	264	259	—	330
Maximum distal width	336	307	352	—	308	352	359	—	478
Width across distal articular surface	208	206	218	—	194	259	254	—	340

* Estimates from figures with allowance for fact that humerus is "less than one-fourth of the natural size."

† No assurance that this is from the same individual as the humerus that bears the same number.

** Measurements supplied by R. A. Stirton and Owen J. Poe.

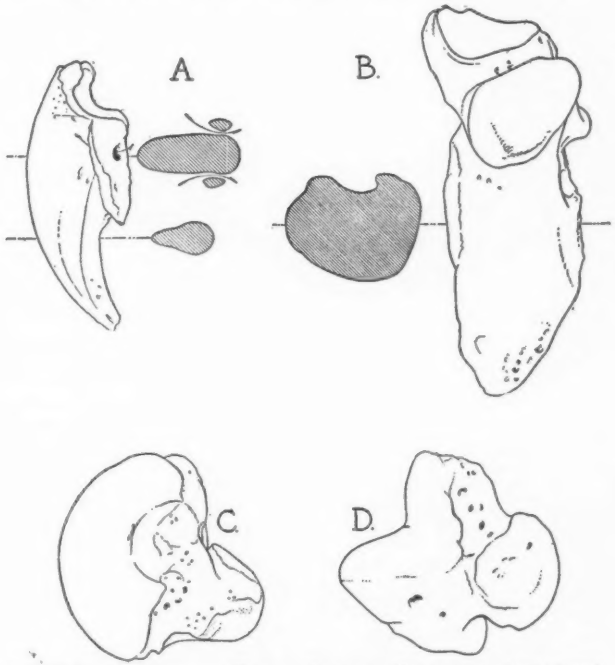

FIGURE 9.—MEGATHERIUM UNGUAL PHALANX, CALCANEUM, ASTRAGALUS

(A) ungual phalanx of third digit of manus, laterial view and cross sections, UC 36941; (B) calcaneum, dorsal view and cross section, UC 36901; (C-D) astragalus, dorsal and lateral views, UC 36907. Hormiguero fossils × ⅛, late Pleistocene, loc. V4201, El Salvador, Central America.

on this continent disclose megalonychid and mylodontid affinities.

Bryan Patterson (personal communication) writes:

"Mention is made of two characters that may distinguish the Hormiguero population from that represented by the individual described as *M. americanum* by Owen. The claw, which seems to me to be manus, III, 3, is said to be 'large and flattened'.

The flattening is an artifact caused by loss of the bony sheath (Fig. 11). When this sheath comes off, as it is very liable to, there is *absolutely* no trace on the dorsal surface that it was ever present. This is very evident in one of our Honduranian specimens in which half of it is missing. The same thing has happened to the type of *Megatherium hudsoni* White.

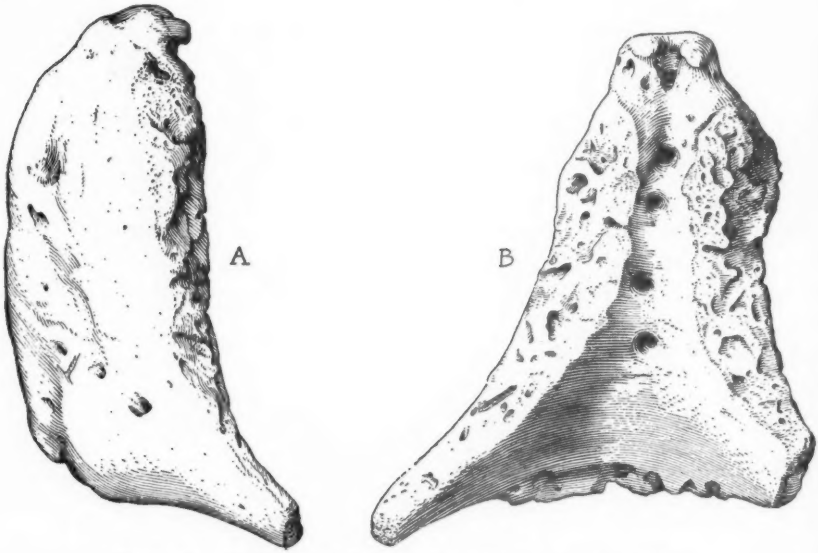

FIGURE 10.—MEGATHERIUM BONE OF QUESTIONABLE IDENTIFICATION

Perhaps developed from separate ossification centers of pelvis at posterior end of pubic bones, (A) lateral, (B) dorsal views, UC 37033. Hormiguero fossils × ½, late Pleistocene, loc. V4201, El Salvador, Central America.

"I strongly doubt the validity of White's proposed species and its supposed Mid-Pliocene age. Available evidence strongly suggests that the Americas were not united until late Pliocene[1], and the Mid-Pliocene Argentinan megatheras are not as large as *hudsoni*. I suspect the specimen of being an accidental inclusion derived from overlying Pleistocene. The type of *hudsoni* is, I believe, manus, III, 3, rather than pes, III, 3, (Fig. 10). The basal portion of pes, III, 3, is short, 60± mm., in our Honduras specimen (cf. Owen, 1859, pl. 41), whereas the corresponding portion in manus, III, 3 is much larger, 110± mm. in ours, 142 mm. in the Hormiguero specimen and 120 mm. (+ ?) in *hudsoni*. White's type and the Hormiguero specimen are both somewhat larger than ours and Owen's (331.5 mm. total length in the Hormiguero claw as compared to 265± mm. in ours), but I doubt if the difference would be significant statistically if a really good series were available. If, as I suspect, the type of '*hudsoni*' is manus, III, 3, White's estimate of the size of his megathere must largely be discounted."

"Our Honduras specimens show that the relative proportions of humeri and femora are the same as in Owen's specimen."

Perhaps the adult humerus and femur from El Salvador are indicative of the intraspecific variation in megatheres. The bones here in question may belong to two animals but we cannot exclude the possibility of their being parts of one skeleton.

According to the Pleistocene records compiled by O. P. Hay, *Megatherium* ranged through the Gulf States and up the Atlantic Coast as far as Mammoth County, New Jersey. In the Mississippi Valley, they have been recorded from Hickory County, Missouri, and in Nebraska; evidently they did not get into the Rocky Mountain or Pacific Coast areas.

CARNIVORA

FELIDAE

The right metacarpal II of a medium-sized *Smilodon* was taken from the Hormiguero quarry

[1] Both megalonychid and mylodontid ground sloths did get through from South America in Hemphillian, Middle Pliocene time, but no megatheres are recorded in those faunas. R. A. S.

(Fig. 4B). These sabre-tooth cats were widely distributed in Pleistocene time.

NOTOUNGULATA

TOXODONTIDAE

A lower M_2 (Fig. 4C) found in the Hormiguero quarry closely approaches those in *Toxodon*, but Mr. Bryan Patterson has informed us that, so far as he can judge from a photograph, a skull from a bog in Costa Rica definitely is not *Toxodon*. El Salvador is perhaps the farthest north these animals have been recorded. The family first appears in the Oligocene of South America, and essentially all of its evolution took place on that continent. Some individuals spread into Central America after the elevation of the marine barrier that separated the two continents during most of the Cenozoic.

PROBOSCIDEA

GOMPHOTHERIIDAE

Several specimens in the Museo Nacional in San Salvador are recorded as found on the coast near La Union. Some incomplete mandibles, a palate, and an ?upper tusk are here referred to the genus *Cuvieronius*.

The tusk fragment is 403.2 mm. long, moderately curved with a proximal diameter of 151.2 x 126.0 mm., and a distal diameter of 113.4 mm., enamel band 66.4 mm. wide and gently spiraled. A palate with M^2 heavily worn, and an M^3 partly worn anteriorly, width between inner anterior edges of M^2 50.2 mm., width between M^2 75.6 mm.; M^3 with 4 lophs, single trefoils extending anteriorly and posteriorly from large lingual cusp, 3 cuspules on labial side of loph, dimensions of tooth-length 132.5 mm., width 93.0 mm., height of lophid above enamel line 67.8 mm. M^2 with 3 well-worn lophs, dimensions of tooth-length 107.1 mm., width 88.6 mm.; cingulumlike structure between lower labial edges of lophids. Mandibles without lower tusk, length from posterior edge of symphysis to posterior edge of M_3 308.7 mm., $4\frac{1}{2}$ lophids on M_3, single trefoils developed anteriorly and posteriorly from labial cusp, 2 inner cuspules on lophid, 3 triangular cusps on 4th lophid, 5th lophid with 2 cusps, dimensions of tooth-length 176.4 mm., width 88.1 mm., height 61.6 mm. M_2 missing but the alveolus showed that the tooth was approximately 94.6 mm. long. There were no other teeth in the jaw. The depth of the mandibular ramus below the anterior edge of M_3 is 150.8 mm.

Another mandibular fragment in the Museo Nacional and labeled as coming from Barranca El Sisimico, Departemento de San Vicente, has an M_3 and part of an M_2 in place. The M_3 is too poorly preserved to show any character. The posterior end of M_3 is not exposed, but the 4 anterior lophids are well preserved, lophids narrower and valleys wider than in the La Union specimens, otherwise the specimens are much alike. The Sisimico specimen is slightly crushed. This specimen is also referred to *Cuvieronius*. We have not been able to make a generic identification for the Hormiguero patella and fibula.

The earliest records of these spiral-tusked mastodons are in the Mount Eden Middle Pliocene of California. If Osborn's identifications and age determinations of cheek-teeth are correct, there are several occurrences in the late Pliocene of the United States and Mexico. Species of the genus may have reached South America also in the late Pliocene since they are common there in Pleistocene deposits throughout the Andes.

ELEPHANTIDAE

There are also remains of mammoth in the Museo Nacional; these bones and teeth were found in association with ground sloth in the bank of the Rio Jerusalen, Hacienda San Lorenzo, Jurisdiccion de Nueva Granada, Departemento de Usulutan (we are not able to locate this site on the known maps or in the field). The relatively narrow transverse plates in the cheek-teeth indicate a close relationship with *Mammuthus jeffersonii* (Osborn).

This is the southernmost record of mammoths in North America and may be close to the southern point of their distribution in the Western Hemisphere. Osborn (1929; 1942) has described two cheek-tooth fragments from South America as *Parelephas columbi cayennensis*. These were "collected by Captain Perret in Cayenne (French Guiana), South America, and now preserved in the Museum d'Histoire Naturelle, Marseille, France, as No. 8449 (cast Amer. Mus. 21933)" (Osborn, 1929). There is no information on the stratigraphy, nor is there a statement as to the nature of the discovery. Possibly these fragments were taken to the port of Cayenne from North America or Europe by someone interested in such objects and eventually discarded. If the mammoths reached South America, it is a most curious fact that no remains have been discovered in the hundreds of Pleistocene vertebrate localities from which specimens have been taken and observed on that continent. Mastodons are common there as are both mastodons and mammoths in North America. Consequently, we hesitate to accept the Cayenne specimen as evidence

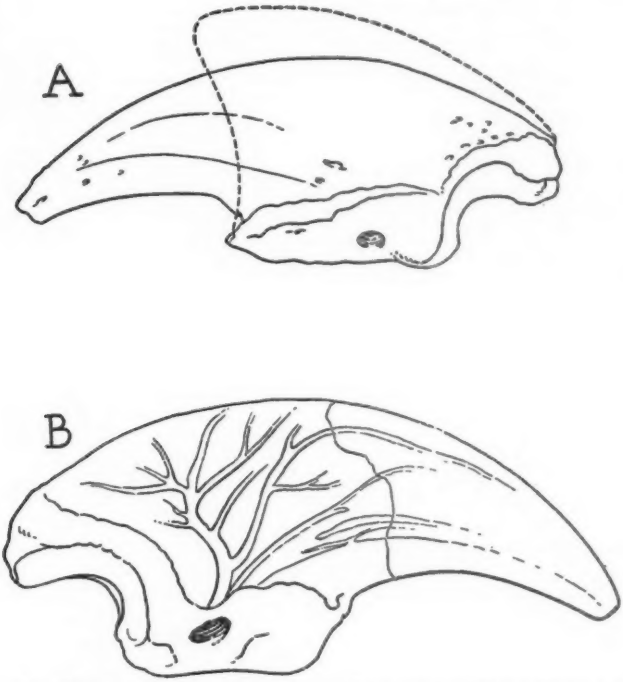

FIGURE 11.—COMPARATIVE FIGURES OF UNGUAL PHALANGES OF THIRD DIGIT OF MANUS

Laterial views × ¼. (A) *Megatherium*, Hormiguero, with dotted lines indicating bony sheath as outlined by Bryan Patterson, UC 36941; (B) *Megatherium hudsoni* White, Florida, MCZ 3694.

for the dispersal of mammoths into South America until it is supported by other discoveries.

ARTIODACTYLA

CAMELIDAE

The only camelid specimen seen in El Salvador was a poorly preserved, small cervical vertebra that was accidently destroyed while we were removing a large ground sloth bone from the Hormiguero quarry.

BOVIDAE

An isolated upper molar (Fig. 4A) of a *Bison* was uncovered in the bottom of the Hormiguero quarry. Bisons may have ranged as far south as Costa Rica. The fossil tracks of some large artiodactyl, possibly the *Bison*, have been recorded in volcanic materials near Managua in Nicaragua.

It might be argued that the presence of *Bison* indicates a much more temperate climate than that in the area today. With the advance and retreat of the great ice sheets on the North American continent, it is logical that there was a much greater extension of the oak-pine association (Arid Upper Tropical of Dickey and van Rossem, 1938) in Central America and Mexico, and this would probably be the bisons' zonal preference, but during the accumulation of the Hormiguero materials, we think that the Central Valley in El Salvador was very little different from the climate there in early historical time. If the oak-pine association had occupied the interior valley and most of the coastal area of El Salvador at that time, more bison, camel, and horse remains might be expected in the assemblage.

Though bisons undoubtedly came from Eurasia via the Bering land bridge, we cannot be certain of the climatic conditions on the bridge at the time of their spread through that area, though we feel safe in assuming it was not tropical; nor can we assume that the bisons that finally reached the lowermost latitudes in their range possessed identical temper-

ature tolerances with their predecessors that spread across the Bering route. One tooth like the one from Hormiguero might be referable to nearly any species in the genus and would be extremely meager evidence in any climatic interpretation.

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Manuscript Received by the Secretary of the Society, November 28, 1947.

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 60, PP. 1755–1770, 6 FIGS. NOVEMBER 1949

FREQUENCY AND PERIODICITY OF OROGENETIC MOVEMENTS

By L. M. R. Rutten

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Abstract

Stille's theory of short, worldwide, and synchronous orogenetic phases, alternating with long anorogenetic periods, is discussed. The facts are not in accordance with this theory. Orogenetic periods occur irregularly, at quite different dates in different regions. Moreover an orogenetic period is not of a very short duration. Much of the evidence used by Stille is unreliable, consisting of inexact or uncritical age determinations.

Introduction

General

It has long been known that for a given region orogenetic and anorogenetic periods alternate. Many geologists have considered these alternations to be worldwide and synchronous. It will suffice here to refer to Marcel Bertrand and Chamberlin and to such expressions as, "Variscian" and "Alpine" orogenics.

J. Joly (1925), in his fascinating book "Surface History of the Earth" tried to explain the relation between the periodicity of orogenics and the great transgressions and regressions of the sea by geophysical data and suppositions. On an absolute time scale, the periodicity of orogenics, as known to Joly (Fig. 1) is not very convincing.

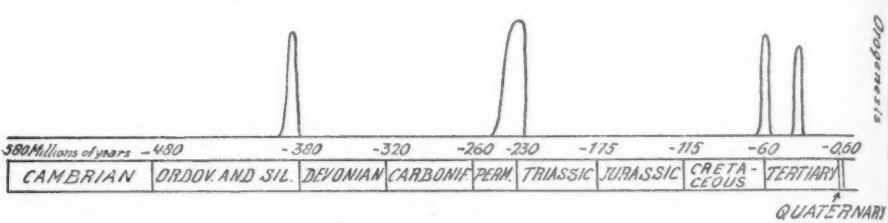

FIGURE 1.—DISTRIBUTION OF OROGENETIC PERIODS, ACCORDING TO JOLY (1925)

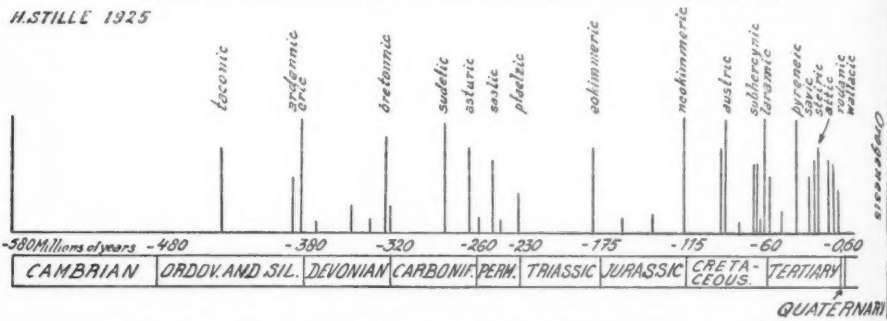

FIGURE 2.—DISTRIBUTION OF OROGENETIC PERIODS, ACCORDING TO STILLE (1924)

Shortly before Joly, Stille (1924) put forward a system of the history of the earth, consisting of an alternation of anorogenic periods of long duration, with short orogenetic phases (Fig. 2). This theory was reiterated in 1940, in an important treatise on the geology of the Americas. In contrast with the ideas of Joly, this theory does not state that the alternation of anorogenetic and orogenetic movements is strictly rythmic. Nor does Stille to my knowledge try to explain the phenomena.

He is however strongly convinced of the synchronism of tectonic movements the world over, and moreover that the anorogenetic periods were of long duration, as against the short orogenetically active phases.

"All mountain building is confined to a relatively small number of narrowly defined phases of ... worldwide importance.... It is found simultaneously in most widely separated regions of the earth." (Stille, 1924, p. 44, 45). This same idea is summarized in Stille, 1940, fig. 1, which is reproduced here (Fig. 3).

Objections to Stille's theory

The orogenetic phases as defined and baptised by Stille have been used by many other authors, not only in Germany, but also in Russia, Asia, China, Scandinavia, and Holland. These authors, too, regard these phases as well established realities, which have occurred simultaneously in different parts of the world.

A number of facts will be enumerated, bearing on these points. They will prove that Stille's theory is untenable. I have chosen these facts from the treatises of Stille and other geologists of his school. In this way, no bias against the theory could be involved in selecting the ma-

terial. Although no complete record can be achieved, it will clearly emerge that the theory is based on a foundation of uncritical use of stratigraphical and paleontological facts.

of the original phases into subphases results in a less marked contrast between long anorogenetic and short orogenetic periods.

Particularly significant is the fact that the

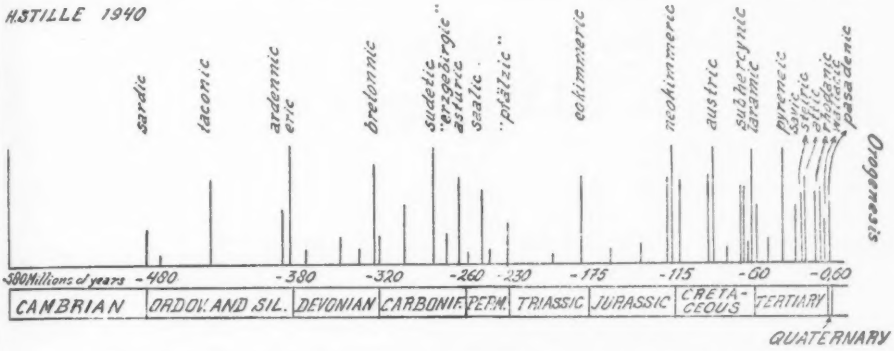

FIGURE 3.—DISTRIBUTION OF OROGENETIC PERIODS, ACCORDING TO STILLE (1940)
The Sardic, Erzgebirgic, and Pasadenic have been added as major orogenetic phases.

I fully realise that a single paper will not destroy the building Stille and his followers have erected. But I do hope that the reader will be induced to make a critical survey of the facts in those regions which are familiar to him. I am convinced that such critical examination will satisfy him of the unsound foundations of Stille's theory. I hope that through this means a better understanding of the history of the earth's crust may be reached.

Historical remarks

At the end of the nineteenth century, the existence of four important post-Cambrian orogenetic periods was known: the Caledonian, Variscian, Laramide, and Alpine periods. The general opinion was that the intervening periods were more or less a-tectonic. The nonconformities related to the four major orogenics are very impressive, and "minor" nonconformities were overlooked for the time being.

Since then new orogenetic periods have been described every time our knowledge of the geology of certain areas increased. In addition, all the old established "phases" have been divided into subphases. The time needed for the major orogenics was found to be considerably longer than was formerly supposed. Every addition of an orogenetic phase or subdivision

younger the period, the more difficult it becomes to discern separate, world-wide synchronic phases. It is these younger layers whose history is best known. Consequently, the more we know of a period, the less Stille's theory becomes applicable.

Moreover, the difference between orogenetic and epeirogenetic movements has not always been clearly drawn. Stille assumes that epeirogenetic movements were going on almost all the time. The theory of short, distinct phases of tectonical activity is expressly limited to orogenetic movements (Stille, 1924, p. 10–15). Epeirogenetic movements occurring together with an orogenetic phase—"which proves that they are a shortlived phenomenon"—do, however, belong to the system of orogenetic phases. Contrarily, epeirogenetic movements in anorogenetic periods may locally, in weak parts of the crust, attain orogenetic aspects (Stille, 1924, p. 17).

This of course, leaves much scope to admit or to reject epeirogenetic or smaller orogenetic movements as instances of a particular orogenetic phase. In practice even the slightest of epeirogenetic movements have been used as an indication for the world-wide occurrence of certain orogenetic phases. This has been a source of much unreliable correlation.

Critical analysis of many correlations and

age determinations used by Stille and his followers as a basis for their theoretical edifice shows, however, that most mistakes are made by accepting correlations not supported by factual data. Commonly, the age determination of a certain movement, is given as: "Either phase a, or phase b." Normally the time interval between the youngest beds folded and the oldest nonconformous layers is rather large. The possibility is never discussed, however, that this movement might have occurred at any time in between these particular phases. I believe that this practice is chiefly responsible for the wide circle of followers which the theory of Stille has found. The interpretation of geological field work is of course much easier, if correlation with only a small, definite number of orogenetic phases is permitted, and it is not necessary to study all the possibilities of an intermediate age.

Orogenetic Phases

Movements from late Algonkian to Ordovician

Between Algonkian and Cambrian.—A very widespread disconformity[1] occurs between Algonkian and Cambrian. Its importance is small however in many areas. On the other hand, there are large regions where the Upper Algonkian passes without a disconformity into the Cambrian. The great neo-Algonkian nonconformity certainly lies much deeper than these unimportant interruptions between Algonkian and Cambrian. Presumably it occurred 100 millions of years or more before the beginning of the Cambrian.

Salair, Sardic, and Trysil phases.—For a long while, Cambrian times have been considered as exceedingly quiet. In his first synthesis, Stille (1924) does not mention any Cambrian movement. Since then our knowledge has been increased and Cambrian movements have now been referred to three different phases. The oldest one is the Salair phase, of pre-Upper Cambrian age, occurring mainly in Siberia. The Sardic phase, occurring during the end of the Cambrian period, is the most important one. Tectonic movements found in Western Europe, in North America, and probably in the Argentine are referred to this phase. I do not feel certain however, that these movements are synchronous everywhere. The third period is formed by the Trysil phase, found in Norway and England, which occurred between Cambrian and Ordovician.

Table 1.—*Movements from Late Algonkian to Ordovician*

Between Cambrian and Ordovician
- Trysil
  - *Age:* Between Cambrian and Ordovician
  - *Europe:* Norway, England (Bubnoff, 1941)

Cambrian
- Sardic
  - *Age:* End of Cambrian
  - *Europe:* Huelva, Spain (Lotze, 1943). Novaja Zemlja (Backlund, 1940). Poland, Sudetes (Bubnoff, 1941)
  - *North America:* Great Acadia, Appalachian Piedmonts (Stille, 1940, p. 36, 37)
  - *Argentine* (?): (Stille, 1940, p. 56)
  - Movements probably not synchronous in all localities
- Salair
  - *Age:* pre-upper Cambrian
  - *Siberia:* Jennesei region; Kusneski Alatau near Salair (Leuchs, 1935, p. 121, 130, 132, 136, 153)
  - *Southern Ural:* (Polutow, 1940, p. 276)

Between Algonkian and Cambrian
- *Widespread disconformities:* South Africa, North Europe, Greenland, Grand Canyon, etc. Movements often unimportant
- *Strict conformity in wide regions:* Northwest America (Stille, 1940, p. 57), China (Lee, 1939, p. 42, 80)

It follows that during the Upper Cambrian there has been tectonic unrest in many places, all over the world. The movements were however of a gentle nature. The perimeter of most areas where Cambrian movements have been detected is rather limited, owing to the enormous stretches of post-Cambrian rocks between the exposures of Cambrian age. We must be very prudent in trying to assign the Upper Cambrian tectonical unrest into three narrowly limited orogenetic periods. Evidence for the synchronous occurrence of the Cambrian phases in different parts of the world is certainly insufficient.

[1] The words disconformity and nonconformity are used differently by European and American writers. Following Moore (1949, f. 22, B, D), the word disconformity is used for a stratigraphical gap, or overlap, or a similar unconformity. A nonconformity is a tectonical or angular unconformity.

Movements during the Ordovician and the Silurian (Caledonian movements)

We have already considered the Trysil phase, occurring at the dawn of the Ordovician. Stille (1924) has distinguished three phases of Caledonian movements. These were the Taconic phase between the Ordovician and the Silurian, the Ardennic phase before the Downtonian, and the Eric phase of pre-Old Red age.

Neosilurian movements.—The Taconic and Ardennic phases are readily discernible; they are separated by a long time interval. Discrimination between the Ardennic and Eric phases is very difficult, however, as they are only separated by the Downtonian (Bubnoff, 1941, p. 59). According to several geologists moreover, the Downtonian forms a distinctive facies only, and not a stratigraphical period. In this case the distinction between Ardennic and Eric becomes illusory.

In Belgium and North France, for instance, it is almost impossible to give a precise definition of the term Ardennic, although this zone includes the type locality. There are many instances of Neosilurian orogenics, but the folding cannot be exactly dated. Nearby, in Great Britain, other phases of the Caledonian movements have been reported, such as Taconic and Eric movements in Northwest Ireland and in Wales. Consequently the orogenics in the Ardennes may also belong to several phases of the Caledonian movements. The terms "Ardennic" and "Eric" ought to be suppressed, because we are only able to distinguish movements

TABLE 2.—*Movements during the Ordovician and the Silurian (Caledonian movements)*

Silurian		
Eric *Age:* pre-Old Red Ardennic *Age:* pre-Downtonian	Neosilurian movements *Age:* between Ludlow and Old Red *Europe:* Many movements in Great Britain; Ireland; along the Ardennes-Poland belt; Scandinavia (Stille, 1924; Bubnoff, 1941) *North America:* Alaska, Northern Appalachians (Stille, 1940) *Asia:* Kusnezki Plateau (Leuchs, 1937), Tscherskij Mts. (Gundlach, 1942) *Australia:* New South Wales; Victoria and Tasmania: great intrusions probably connected with Neosilurian movements	Caledonian movements *Age:* between Ordovician and Devonian *Europe:* Various movements in Great Britain; Scandinavia and South and Central Europe. Movements variously dated from Taconic to Eric in Spitzbergen. Traces of movements in Greenland *Africa:* South Sahara (Stille, 1924) *South America:* Probably existing *Asia:* Russian Altai; Eastern Siberia; Sajan Territory (Leuchs, 1935, p. 160, 173; 1937, p. 52). Kara-Tan, Chu-Ili Ranges, North Tien San, Northeast Kzahkstan (Norin, 1940)
Between Ordovician and Silurian Taconic *North America:* Appalachians, type locality between Massachusetts and Gaspé Peninsula (Stille, 1940, p. 37, 38) *Europe:* South Wales, Shropshire, North Ireland (Stille, 1924, p. 66, 67) *Australia:* Victoria, Tasmania (Andrews, 1938). *Asia:* Wilni; Kirchizes (Leuchs, 1937). Tscherskij Mts. (Gundlach, 1942)		
Ordovician Ekne *Age:* Upper Ordovician Trondheim *Age:* Middle Ordovician *North America:* Cincinnati axis (Bubnoff, 1941) Trysil (see table 1)	*Europe:* Norway (Bubnoff, 1941, p. 163, fig. 76)	

belonging to a Neosilurian phase. No definite age can be given for these orogenics.

The Neosilurian movements are widespread (Table 2). The Neosilurian movements however, took a conisderably longer time than did, for instance, the Taconic movements.

TABLE 3.—*Movements during the Devonian*

Devonian *Greenland:* Four orogenetic phases between upper middle Devonian and uppermost Devonian (Stille, 1940, p. 54) *Asia:* Late-Devonian folds, Kusnezki Alatau. Post-middle Devonian folds, Russian Altai. Frequent Devonian folding in the Khirghizes (Leuchs, 1935, p. 155–157, p. 160; 1937, p. 28) *Australia:* Unconformities between middle and upper Devonian and at the end of Devonian in Victoria and New South Wales. The latter might already be Bretonic (Andrews, 1938) *Europe:* Disconformity in Old Red of North England and Scotland (Stille, 1924, p. 78) *Europe:* Strong movements between Silurian and upper Coblencian. Probably a continuation of Caledonian movements. (Stille, 1924, p. 78–80)	*North America:* Epirogenetic movements in the central U. S. A. and orogenetic movements in the Appalachians.

Other Caledonian phases.—Lately two new phases have been added to the Caledonian movements in Northwest Europe. These are the Trondheim phase of the Middle Ordovician and the Ekne phase of the Upper Ordovician (Bubnoff, 1941, p. 163). A system has thus been devised for Northwest Europe, consisting of no less than five different orogenetic phases for the Ordovician and Silurian. It goes without saying that it will be difficult, at best, to correlate the orogentic phases found outside Europe with this system. The Trondheim and Ekne phases are almost unknown outside their type localities in Norway. Bubnoff (1941) mentions the Cincinnati axis from North America, but this is only a very gentle uplift.

Many movements have been assigned to the Taconic phase, but it has not been proved that all these movements occurred simultaneously between the Ordovician and the Silurian. Besides the well-dated movements assigned to the Taconic or Neosilurian phases, many others have been found, which can only be described as "Caledonian". An incomplete summary of these movements has been given in Table 2.

Movements during the Devonian

Without doubt the Devonian forms a quiet period. However, the duration of the Devonian is only half that of Silurian and Ordovician combined, and equally only half that of the Cambrian. One is always inclined to regard the Devonian as a long period, as in Europe and North America sediments of this formation are so very important. Still, Devonian movements are few and quantitatively unimportant. Moreover they are widely scattered and not found in great continuous belts.

Such movements as have been reported, are very variable as to age (Table 3). If desired, one could distinguish some 7 or even 14 different orogenetic phases for this period. Consequently, it is almost impossible to correlate the various Devonian movements in different parts of the world.

Movements during the Carboniferous and the Permian (Hercynian or Variscian movements)

Introduction.—Contrasting with the Devonian, the Carboniferous, and the Permian have been periods of enormous tectonic activity. This begins with an orogenetic period at the beginning of the Carboniferous, called the Bretonic phase. Some authors now even divide this phase into three subphases, the Marsic, Nassauic, and Selkic. These are followed by three phases in the Carboniferous proper, the Sudetic, Erzgebirgic, and Asturic. Lastly, the Saalic and the Pfälzic have been distinguished for the Permian period. From this enumeration alone, it will be clear that it is often impossible

to correlate Hercynian movements elsewhere with this purely European system.

Bretonic phase.—The movements at the end of the Devonian and the beginning of the Carboniferous form the Bretonic phase (Table 4). The more detailed subdivision has no sense, as it is impossible to correlate these subdivisions even in different parts of Europe. There is no proof whatsoever that the Bretonic movements were strictly synchronous. On the contrary, the Bretonic must have been a phase of long duration, estimated at a couple of million years.

Sudetic phase.—This phase occurred between Lower and Upper Carboniferous. Its type locality is South Germany, where it forms the chief phase of the Variscian orogenesis. It has been found in many other parts of Germany, but even in so small an area, doubts exist as to the strict synchronism of these movements in different parts of the country.

Inexact dating of orogenetic movements.—Stille (1924, p. 89–91) mentions movements in the Alps and in different parts of France, which are "either Sudetic or Asturic". It is not clear why these movements might not have occurred *between* the limited periods of the earth's history, to which were assigned the Sudetic and Asturic phases. These movements might be Erzgebirgic, or they may not be correlated at all to one of the tectonic phases postulated by Stille. As far as the field evidence goes, they may quite possibly belong to intermediate tectonic movements, occurring at some time during the Upper Carboniferous.

This way of arguing is frequently found with Stille and other geologists of his school. They adhere to the hypothesis of a few short and well defined tectonic periods, separated by times of tranquility. In studying a certain movement, they only care if it belongs to one or the other of a certain pair of phases. The possibility that the movement might have occurred between these phases, or that it might belong to an intermediate, but hitherto unrecognized phase, is put aside beforehand. This tends to slur the critical dating of tectonic movements. It is of course much easier to correlate a certain movement with one definite phase out of a short list of periods of tectonic activity, than to prove that it did not occur at some time during the intervening period. *Every correlation whereby tectonic movements are simply correlated to a certain tectonic phase, must be judged very critically in the light of these considerations.*

Asturic phase.—The next younger orogenetic phase, the Asturic, is of great importance. It occurred at the end of the Westphalian, and its type locality is Asturia, Spain. Here a well defined unconformity has been found in the Basin of Cantabria. It is located within the Upper Carboniferous, supposedly between Westphalian and Stephanian.

Upper Carboniferous unconformities have been found in other basins of Spain, Portugal, and France. I have the impression, however, that their exact age has not been established in any of these cases. The same may be said about the folding of the great coal belt of England, North France, Belgium, Holland, and Northwest Germany. This is generally regarded as having occurred in the Asturian phase, which is probable but not certain. We may only say that the folding is post-upper Westphalian and pre-middle Permian.

The data on the asturian movements are unsatisfactory. Only at the type locality are they well dated; in all other regions their synchronism is far from convincing.

Saalic phase.—Approaching the end of the Paleozoic, it becomes more difficult to date exactly the tectonic movements. The Saalic phase, which derives its name from the river Saale, is said to occur between Autunian and Saxonian (between Upper and Lower Rotliegendes). Stille (1924, p. 100–102) cites nine cases of Saalic movements occurring in Germany. In most cases however, he cannot prove that these movements indeed occurred in so limited a time.

In France an unconformity between Autunian and Saxonian is indicated in the Morvan, the Mont Noir, and near Autun. In Ariège, Asturia, Bretagne, and South England on the contrary, the exact date of the folding is not known. An Upper Paleozoic disconformity occurs, but it cannot be dated as Saalic with certainty. The same may be said of Upper Paleozoic movements from a number of other localities (table 4).

Basic difficulties of correlating orogenetic periods.—Many geologists assume that the chief folding in the Appalachians occurred in Lower Permian times. However, there is still no *com-*

TABLE 4.—*Movements during the Carboniferous and the Permian*

Between Permian and Triassic			
Pfälzic			
No important orogenetic movements *Europe:* Saar-Nahe region, Plateau Central. Perhaps South England and the Ural. Small scale phenomena. (Stille, 1924, p. 105–106). *Australia:* Late Permian orogenesis along East coast (Andrews, 1938)			
Permian			
Saalic *Age:* between Autunian and Saxonian. *Europe:* South Germany; France: Morvan, Mont Noir, Autun (Stille, 1924, p. 100–102) Novaja Zemlja (Frebold, 1943) *Asia:* South China: Tunguru orogenesis (Lee, 1939, p. 149) Upper Paleozoic movements have often been dated as Saalic on insufficient data.	"Asturian" = post-Westphalian and pre-middle Permian. Coalbelt of England-France-Belgium-Holland-Northwest Germany.	Upper Paleozoic movements, probably dating from the lower Permian *Europe:* Ariège, Asturia, Bretagne, South England Alps, Donetz Basin and Ural *North America:* Appalachians. *South America:* Unconformaties between Lower and Upper Permian in the Andes (Stille, 1940, p. 130–139) *Asia:* Asia Minor (Stille 1924, p. 119)	"Hercynian" or "Variscian" movements, not definitely dated. For instance: Large parts of Cordillera of South America and vast territories in Central Asia
Carboniferous			
Asturic *Age:* Upper Westphalian *Europe:* Basin of Cantabria, Asturia, type locality. Also other basins in Spain and Portugal. (Stille, 1924 p. 93, 94) France (Stille, 1924, p. 96–99) Schwarzwald, Zwickau, Sudetes (Stille, 1924, p. 94–96) Synchronism doubtful. Beren Island (?) (Stille, 1924 p. 97, 98. Not proved) *China:* Orogenetic movements in Yunnan and Indochina. Epirogenetic phases elsewhere (Lee, 1939, p. 139)		Movements classified as "either Sudetic or Asturic" may have occurred at any time during the upper Carboniferous For instance: "Midcarbonic" movements in the Eastern Alps, different parts of France and Atlas Mts. (Stille, 1924, p. 89–91, 117)	
Erzgebirgic *Age:* Upper Namurian and lower Westphalian (Kosmatt, 1927) *Europe:* Small area of Erzgebirge (Stille, 1924, p. 100–102) *North America:* Witchita Mts (?). Dating doubtful (Stille, 1940, p. 78–79).			
Sudetic *Age:* Between lower and upper Carboniferous *Europe:* South Germany, type locality. Here it forms the chief Varisician orogenesis. No well dated Sudetic movements outside type area. *North America:* Local small disconformity in Great Acadia (Stille, 1940, p. 67–69)			
Bretonic (Marsic, Nassauic and Selkic subphases cannot be distinguished) *Age:* Lower Mississippian. *Europe:* Laval syncline, type locality. Basin of Ancenis, Loire. Sierra de la Demanda. Sierra de Badajoz. Faults in Scotland. Traces of movements in Westphalia, the Harz, Saxonia, the Sudetes. (Stille, 1924, p. 84–86) *North America:* Wichita Mts (?) (Stille, 1940, p. 77) *Asia:* China (?) (Lee, 1939, p. 125). Tonkin (Fromaget, 1935, p. 169) Russian Altai and Khirghizes (Leuchs, 1935, p. 169; 1937, p. 28)			

munis opinio on this subject. During the last Congress for Carboniferous Stratigraphy at Heerlen in 1935, important differences arose between the paleobotanists Darrah (1937, p. 109–129) and Jongmans (1937, p. 363–392) as to the age of certain formations. This is an example of the ever-present uncertainty of the paleontological synchronization of formations. Everyone knows how many times paleontological specialists fail to agree on certain correlations. As, after all, every dating of specific tectonical movements ultimately rests on paleontological evidence of some kind, this forms a serious drawback.

A second cause of unreliability in Stille's work, already mentioned in an earlier section, is the desire to bring all tectonic movements to a limited number of active tectonical phases. For

instance, during the Variscian orogenetic period, movements may be either Bretonic, Sudetic, Erzgebirgic, Asturic, Saalic or Pfälzic. Folding between these times did not occur according to Stille. *Septimus non datur.* I am trying to convince my readers that real proof of this thesis has never been offered by Stille.

Thirdly, it is always presumed that tectonic movements have been achieved in a relatively short time, thus separating short active tectonic phases from long quiet interludes. No real proof of this assumption has ever been offered either.

Pfälzic phase.—It is generally understood that no important mountain building took place during this last phase of the Hercynian movements.

Movements during the Triassic

Introduction.—For long, geologists thought that during a considerable time before and after the Variscian movements all was very quiet on earth. The Variscian orogenesis consequently stood out not only because of its very important movements, but also by its isolated position. These ideas about the "privileged" position of the Variscian era have suffered badly. Just as the Devonian was not nearly as quiet a period as was supposed formerly, a period of very strong mountain-building has been discovered in some parts of the earth for the Triassic. The apparent quietness of post-Variscian times has hereby been disturbed.

Labinic phase.—These orogenetic movements, called the Labinic phase by Robinson (1937), were effective mainly in Asia, where they have been described from the Caucasus Mountains, from Northeast Asia, China, Japan, and Indochina.

Movements during the Jurassic

Introduction.—During the Jurassic, important movements occurred, which are generally indicated as Kimmeric. The *Eokimmeric phase* at the beginning of the Jurassic and the *Neokimmeric phase*, occurring during the Portlandian or between the Jurassic and the Cretaceous, are commonly distinguished.

Eokimmeric phase.—Even in Europe the facts are not consistent with the hypothesis of a short Eokimmeric tectonic activity at the beginning of the Jurassic. Only in the Crimea, Lias is found disconformably on Norian. Elsewhere, the time interval covering the nonconformity may be larger, and the dating consequently less strict. This is the case, for instance, in Galicia and in Poland, where Dogger is found on Muschelkalk, so that the whole of the Keuper and the Lias are available for these movements. Or else, two separate phases are found, such as in the Donetz Basin.

The data from other continents (Table 6), also indicate a rather variable dating of Eokimmeric movements. For North America they have been described from several localities. But even Stille is not satisfied that all movements during the Jurassic belong either to the Eokimmeric or to the Neokimmeric phases. For a local, feeble nonconformity, situated between Callovian and Malm, he follows Crickmay, in calling this an Agassizian movement (Stille, 1940, p. 188).

In Africa, the folding of the Cape Mountains offers another example of inexact dating. Although proclaimed Eokimmeric, this orogenesis might have taken place at any time during the Middle or Upper Triassic, as the youngest beds

TABLE 5.—*Movements during the Triassic*

Between Triassic and Jurassic Eokimmeric (see Table 6)	
Triassic Labinic *Age:* Between Ladinian and Carnian *Europe:* Caucasus, type locality. (Robinson, 1937) *Asia:* Northeast Asia (Gundlach, 1942), China (Lee, 1939, p. 156) Japan: Major folding of Inner zone of Southwest Japan (Zollich, 1942). Indo China: strongest orogenesis (Fromaget, 1940)	*North America:* Orogenetic activity in West Nevada. Duration very long

folded are the Beaufort Beds of the Lower Triassic.

In Northeast Asia two separate orogenetic periods have also been found, occurring between Norian and Neocomian. They are however not strictly synchronous with the Eokimmeric and Neokimmeric phases. For the older movements the whole time interval from Norian to Dogger is available, whereas the younger occurred sometime between Dogger and uppermost Jurassic.

In China, quite another picture is found; two separate phases are reported, one post-Triassic and pre-Jurassic, the other post-Rhaetic.

Many movements from the Jurassic period have been reported from Siberia, but in most places of the region it is not yet possible to distinguish between Eokimmeric and Neokimmeric phases.

We may conclude that Eokimmeric movements occurred during a fairly long period, extending from Upper Triassic to Middle Jurassic. It seems impossible to find two localities, where exact synchronism could be established for the movements reported from this period.

Neokimmeric phase.—In its type region, the Crimea, the Neokimmeric movements occur between the Jurassic and the Cretaceous. In other regions of Southeast and West Europe, the movements are not strictly synchronous, however, and occur somewhat earlier. In Germany, the chief phase of the Saxonic orogenesis is regarded to be of Neokimmeric age. Here three subphases are distinguished by Stille, the Deister, Osterwald, and Hills subphases. Whether one of these coincides with the movements in the Crimea is not clear.

Neokimmeric movements in other continents (Table 6) are most important in the Pacific region of North America.[2]

[2] It has been brought to my attention that recent work by Taliaferro (1943) has shown that the nonconformity at the base of the Knoxville is nonexistent. Moreover the information Stille gives on British Columbia does not tally with the results of Buddington and Chapin (1929). We are sure that every geologist will be able to add instances of movements which have or have not occurred in contradiction to the rigid system of tectonical phases proposed by Stille. Moreover the general treatises on regional geology will always contain some errors, as their authors themselves are only familiar with part of the regions covered. This paper, however, is intended as a refutation of the ideas on synchronism of orogenetical periods, and to this end Stille's own work is mainly used. Local discrepancies fall outside the scope of this paper.

Movements during the Cretaceous

Austric phase.—Various movements which have occurred during the Lower Cretaceous are assigned to the *Austric phase.* It seems to me that the delimination of this phase is still worse than that of its predecessors. Stille (1924, p. 151) says:

> As to the exact age of the *Austric phase*, in many regions (Karpathians, New Zealand) it is not only pre-Cenomanian, but pre-upper Gault. Elsewhere, as in the Pyrenees, it lies between Upper Gault and Cenomanian. We have to consider two subphases of the Austric, one of pre-Gault, or at least of pre-Upper Gault age, the other occurring between Gault and Cenomanian.

I want to ask the old question again. Why have there been two distinct, sharply separated phases? Why not a longer period of locally intensified tectonic activity? Or why not admit that orogenetic periods are not synchronous in different places on earth? There is no proof whatsoever that the assumption made by Stille (1940, Fig. 1; *see* our Fig. 3) of two prominent, narrowly dated, and well separated phases, is correct.

In fact, the age of the movements brought to the Austric phase is even more variable and may include the Cenomanian and the Lower Turonian. The type locality for these movementes is in the German-Austrian Alps, where the Gosau nonconformity separates Neocomian and Turonian. Intra-Cretaceous movements are known from many other localities in the Alps. (Stille, 1924, p. 145–149). They can however, be only dated as "between Neocomian and Upper Turonian" in most cases. Locally they are pre-Cenomanian, elsewhere they may be post-Cenomanian.

Movements of the Austric phase are known from many localities (Table 7), but their age varies considerably.

Subhercynic phase.—The second tectonical phase recognized from the Cretaceous is the *Subhercynic phase.* In its type locality, along the northern flank of the Harz Mountains, it is very well dated and it can even be divided into two

subphases. The older Ilseder subphase occurred between the Lower Emscherian and the Upper Emscherian. The Werningeroder subphase is "Upper Cretaceous". Moreover, it is probable that the duration of these tectonic movements in individual localities has far exceeded the short

TABLE 6.—*Movements during the Jurassic*

Jurassic

Neokimmeric (Nevadic)

Age: Upper Jurassic and between Jurassic and Cretaceous

Europe: Crimea, type locality (Stille, 1924, p. 131) Caucasus, age between lower Malm and Tithonian. Banaat, age upper Tithonian. Balkan Peninsula, insufficiently dated. Three subphases in Germany, the *Deister*phase of lower Portlandian age, *Osterwald*phase of middle Portlandian age and the *Hills*phase of upper Portlandian age.

Boulonnais, Aquitaine, Ebros Basin, exact ages unknown. (Stille, 1924, p. 131–142)

North America: Important in Pacific region.

California: Nonconformity at base of Knoxville; Intrusion of Sierra Nevada Batholite; nonconformity between Jurassic and lowermost Cretaceous in Klamath Mts. (Stille, 1940, p. 169, 175–177, 181, 182, 185) British Columbia, Coast Range; Yukon; South Alaska. (Stille, 1940, p. 186, 190, 191, 195)

Africa: Western Atlas Mts(?); South Tunis(?) (Stille, 1940, p. 143, 144)

Asia: Tscherkij Mts; Turkmenistan; Pamir; Middle Kwen Lun (Leuchs, 1935, 1937) China (Lee, 1939, p. 175–179)

Agassizian

Age: Between Callovian and Malm

North America: Lower Frazes, type locality (Stille, 1940, p. 183)

Between Triassic and Jurassic

Eokimmeric

Europe: Crimea, type locality. Dobrudscha, East Alps, Galicia, Poland. Two phases in Donetz Basin, between lower and upper Lias, and between Bajocian and Bathonian (Stille, 1924, p. 133–135)

North America: Many localities. Nevada, intra-liassic nonconformity. Blue Mts., Oregon, Lias unconformable on Triassic. Faulting in Appalachians (Stille, 1940, p. 168–170)

Mexico: Zacatecas. Folding of Carnian, unreliably dated. (Stille, 1940, p. 582)

Africa: Cape Mts., folding between lower Triassic Beaufort Beds and lower Liassic

Asia: Northeast Asia, folding between Norian and Dogger (Gundlach, 1942) China, two phases, post-Triassic/pre-Jurassic, and post-Rhaetic. (Lee, 1939, p. 168–170). Tscherkij Mts. (Leuchs, 1935, p. 177) Transkaspia (Leuchs 1937, p. 11)

Asia: Movements during the Jurassic, insufficiently known to be dated exactly. Siberia; Southeast Gobi and Ordos; East Tien Shan; Kwen Lun (Leuchs, 181, 1937, p. 11, 70, 92, 122, 197)

found between the Heidelberg beds and the Heimburg beds, which together form the Granulata beds, this being a local name for the lower part of the Lower Senonian.

Tectonic movements are reported from the Upper Cretaceous in many places in Europe, the Antilles, and South America (Table 7). They are not strictly synchronous with those from the northern Harz Mountains, even in Europe. In most cases, their age can only be ascertained as time periods available for the Ilseder and Werningeroder subphases.

According to Stille (1940, p. 580–584), the Subhercynic phase forms the maximum orogenetic period in South America. In 1924 however, extensive Austric movements were reported from Cubut and from southernmost South America (Stille, 1924, p. 151). Moreover, the chief orogenesis was then reported as being of Laramic age (p. 156). As our knowledge of

the geology of South America has not increased vastly during this period, the shifting of orogenetic periods seems to be mainly the result of a different interpretation by Stille. The reasons for this shifting of orogenetic periods are not clearly discussed. It seems preposterous to me that, while shifting orogenetic periods from Austric to Subhercynic and from Laramic to Subhercynic, involving periods of 25 millions and 15 millions of years, an author still postulates that in between these "definite" tectonic phases no orogenetic movements have occurred. There is no reason why the folding in South America did not take place at one or more intermediate periods between the Austric, Subhercynic, and Laramic phases.

TABLE 7.—*Movements during the Cretaceous*

Cretaceous
- Subhercynic
  - *Age:* Upper Cretaceous
    - ***Werningeroder subphase***
      - *Age:* Lower Senonian
    - ***Ilseder subphase***
      - *Age:* Middle Emscherian
    - *Europe:* Harz Mts., type locality. North Germany, French-Lombardic- and Julian Alps, Dibra region in Caucasus Mts. Movements here occurred sometimes during the upper Cretaceous. No strict synchronism. (Stille, 1924, p. 151–154)
    - *Antilles:* Jamaica, Cuba, Haiti, Bonaire, Curacao (Stille, 1940, p. 338, 345–347, 356)
    - *South America:* Guaduas; Guadeloupe; Northeast Columbia (Stille, 1940, p. 388, 392) Transgression of red formation along the Andes mostly upper Cretaceous, "Subhercynic", possibly lower Tertiary "Lamarid", in part. (Stille, 1940, p. 470, 477)
- Austric
  - *Age:* Badly dated, variable between lower Gault and Turonian.
    - *Europe:* Gosau, type locality, and many other localities in the Alps. Pre-Cenomanian age locally proved, elsewhere age between Neocomian and upper Turonian (Stille, 1924, p. 145–149) Pyrenees, age between upper Gault and Cenomanian (Stille, 1924, p. 151) Swiss Jura; South Yougoslavia; Greece; Karpathians.
    - *North America:* Pacific region. Intrusions and tectonic manifestations in Cretaceous. Badly dated. Alaska; Canada, Mexico (?) (Stille, 1940, p. 195, 256, 304, 305).
    - *Asia:* Hindukush, intracretaceous activity (Stille, 1924, p. 149). Northeast Asia, important *Kolymic phase*, of Neocomian or Albian age (Gundlach, 1942). Sachalin, age Valanginian-upper Albian. Japan, Austric movements important, varying in age from pre-upper Albian upwards (Zöllich, 1942)
    - *New Zealand:* (?) Albian nonconformable over older Cretaceous.

Movements during the Tertiary

Introduction.—Much the same facts and arguments relate to tertiary movements. No new viewpoints are added and it is only for the sake of completeness that I include a chapter on Tertiary movements.

Laramic phase.—The oldest phase concerned is the Laramic phase occurring between the Cretaceous and the Tertiary. As we have seen the Subhercynic phase tends to spread out over a considerable portion of the Upper Cretaceous. Consequently it will be often difficult to distinguish between Laramic movements and the younger of the Subhercynic movements. It is well known that locally the paleontological separation of uppermost Cretaceous from lowermost Tertiary may be very difficult indeed.

Type region for the Laramic phase is the Rocky Mountains, where the chief orogenesis took place during this period. In the zone from Alaska to Mexico, two subphases are known from many places, one dating from the latest Cretaceous, the other from the upper Paleocene (Stille, 1940, p. 256, 257). But, although a great amount of detail is now known from western North America, it remains difficult to correlate these two subphases with lower Tertiary movements elsewhere.

Character of younger orogenetic phases.—Out of the different phases proposed for the younger

Neozoicum, the Pyreneic phase and the Pasadenic phase are acceptable as more or less separate entities, comparable with the phases from the older periods. They may be sharply delimited orogenetic periods, as postulated by Stille, or we may use these indications as a general time-index, as I have retained the names of the older phases. But I do sincerely believe that it is absolutely impossible to correlate and to distinguish between the *Savic, Steiric, Attic, Rhodanic,* and *Wallachic phases* in different continents.

As I have already indicated, it becomes increasingly difficult to date, correlate, and distinguish separate phases of tectonic activity, the younger they are. It becomes harder and harder to accept the postulate that these separate periods of tectonic activity have any real significance.

TABLE 8.—*Movements during the Neozoicum*

Quaternary

Pasadenic

Age: Pleistocene

California, Siwaliks, Java, not covered in this paper.

Tertiary

Wallachic, Rhodanic, Attic } Age: Pliocene!

Wallachic — *Type locality:* Wallachia, South Karpathians.

Rhodanic — *Type locality:* Rhône region, France.

Attic — *Type localty:* Attica, Greece.

Post-Pyreneic movements
Known from many parts of the earth. Age determination and correlation very hazardous

Steiric

Age: Middle Miocene. Two subphases!

Europe: Near Graz, type locality (Stille, 1924, p. 185)

Savic

Age: Post-Oligocene/pre-Miocene.

Europe: Save Basin, Yugoslavia, type locality. Age post-Chattian/pre-Aquitanian. (Stille, 1924, p. 178) Important orogenetic period in the Alps

Pyreneic

Age: between Eocene and Oligocene

Europe: South flank of Pyrenees, type locality. Two subphases, one post-Lutetian, the other post-Ludian. Chief orogenesis in Pyrenees. North Spain, Montpellier, France; the Alps; South Europe. (Stille, 1924, p. 165–175)

Asia: Asia Minor (?) (Stille, 1924, p. 175) Formosa, age post-Eocene/pre-Burdigalian (Zöllich, 1942)

The Antilles: (Stille, 1940, p. 350, 379)

Between Cretaceous and Tertiary

Laramic

Age: Uppermost Cretaceous and Lower Tertiary.

North America: Rocky Mts., type locality (Stille, 1940, p. 256, 257). Many reports from Alaska to Mexico and in the Antilles. Two phases, one latest Cretaceous, the other upper Paleocene

Europe: North Spain; Paris Basin; Northern Coal-Belt; French-Eastern- and Southern Alps; Dinarides. Movements not synchronous; some time between upper Cretaceous and Eocene. (Stille, 1924, p. 136–162)

North Africa: (?) (Stille, 1924, p. 162; Krenkel 1925/'37.

Asia: Asia Minor, exact age unknown. (Stille, 1924, p. 163). Anadyr region, Northeast Asia, between Paleocene and Eocene (Gundlach, 1942) Japan, unimportant, evidence not convincing (Zöllich, 1942)

Determining the age of the various layers of the earth becomes more and more difficult, and, contrarily, the time interval separating the different phases grows less and less in younger times. Everywhere the fossil content of the beds varies far more widely with the facies than with the age. Geographic latitude, character of the sea in marine formations, or of the land in continental formations, are of the greatest influence on fossil assemblages. Consequently, long-distance age correlations are very difficult. This is evidenced by the fact that the age determinations of many tertiary formations have under-

gone frequent changes. (See, for instance, Senn, 1940.)

Some data for these younger orogenetic periods have been compiled in Table 8. Even to Stille, who postulated the existence of these

Conclusions

Alternation of orogenetic and anorogenetic periods for limited areas of the earth's crust is an established fact.

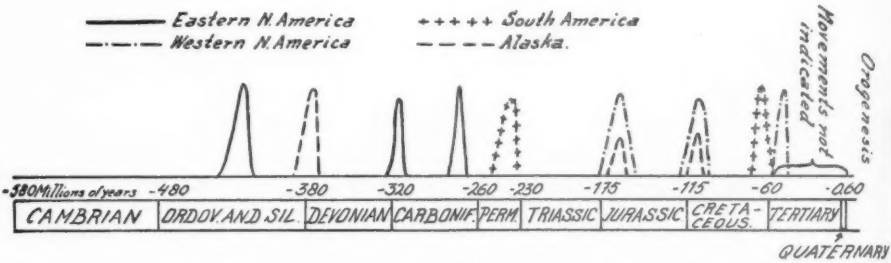

FIGURE 4.—DISTRIBUTION OF OROGENETIC PHASES IN ALASKA, EASTERN NORTH AMERICA, WESTERN NORTH AMERICA AND SOUTH AMERICA. COMPILED FROM STILLE (1940)

phases as separate tectonic movements, the difficulties are apparent sometimes. For instance, the Savic phase is defined as being post-Oligocene and pre-Miocene. Its age can be more narrowly fixed as post-Chattian and pre-Aquitanian. But these beds can only be separated with difficulty, especially in continental facies. For Tertiary basins outside Europe, an exact correlation with the European time scale is impossible. The solution of the difficulty is however obvious to Stille, who says: "In such cases the position (of the beds) in relation to the Savic disconformity, may support the age determination" (Stille, 1924, p. 178). This is putting the cart before the horse!

For the short Pliocene period, Stille has created three different orogenetic phases, the Attic, Rhodanic and Wallachic phases. It is quite impossible to say which movements in Asia, Africa, or America should correspond to the movements of these "phases" in their type localities in Greece, the Lower Rhone Valley, and Northern Roumania. It is moreover extremely improbable that these movements are everywhere synchronous with these three "short phases of orogenetic activity," and separated by "long periods of orogenetic tranquility." This is however postulated by Stille (1940, Fig. 1; Fig. 3 of this paper). It must be clearly evident that the designation of separate orogenetic phases has here been pushed *ad absurdum*.

In most cases, the anorogenetic periods are long, the orogenetic periods short. We may therefore speak of orogenetic revolutions, or orogenetic phases.

A correlation in time for orogenetic movements in widely separated regions of the earth is often found. This correlation remains however of a very general character. No correlation of sharply limited orogenetic phases can be established over wide areas.

Moreover, the incidence of orogenics of about the same age in different areas is certainly not higher than that of orogenetic phases occurring only in one area or orogenetic belt. As an example, the major orogenetic periods in the Americas (Stille, 1940), have been compiled in Figure 4. Although several orogenetic periods do occur in two areas more or less simultaneously, many others are only found in a single region.

Correct dating and correlation of tectonic movements is very difficult, and must depend on paleontological and stratigraphical facts. The dating of many orogenetic movements, as accepted by Stille for a basis to his theory, does not stand a critical scrutiny of the facts available.

The wider the area under consideration, the greater the difficulties of effective correlation. Consequently, the more widely separated certain orogenics are, the vaguer their age correla-

tion, and the vaguer are the conclusions which can be drawn as to their simultaneous occurrence.

Moreover, the orogenics are not confined to a single short burst of orogenetic activity. On the contrary, several active phases of mountain building occur during a rather long period of general tectonic unrest. These movements together form the mountain-building period. The real state of affairs, even in limited areas, is not a strict alternation of quiet anorogenetic time and orogenetic revolutions. For regional geological history, the picture given by Lee (1939) is much more according to fact than those of Stille (Figs. 1, 2, 5).

If we want to draw a graph of the incidence of tectonic activity during geological history the world over, Figure 6 (Umbgrove, 1942, Table II) is again much better adapted to the facts than Stille's graphs. Superposed on a general tectonic unrest, persisting throughout the history, various periods are characterized by more pronounced orogenetic movements. These periods are not worldwide however, nor are they sharply delimited; neither do they contrast sharply with the intervening periods of lesser tectonic activity.

Instead of worldwide, synchronic, orogenetic revolutions, there thus have been periods of long duration, characterized by varying and fluctuating tectonic activity. The active periods are not worldwide. While part of the earth was in tectonic rest, elsewhere tectonic activity was found. The quiet regions may have already been folded during an earlier date, or they will be folded at a later date, or they may remain ultimately undisturbed. Differences in time of folding may be found not only in different continents, but also along one and the same orogenetic belt.

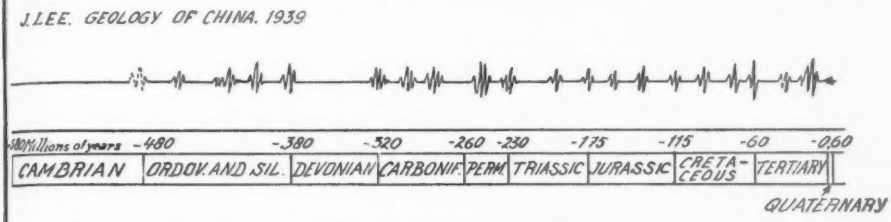

FIGURE 5.—DISTRIBUTION OF OROGENETIC PERIODS IN CHINA, ACCORDING TO LEE (1939)

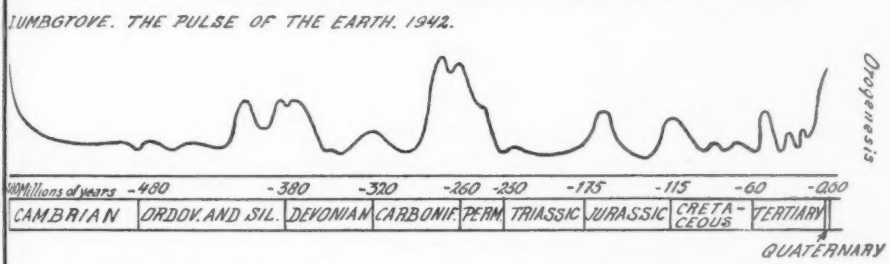

FIGURE 6.—DISTRIBUTION OF TECTONIC ACTIVITY ON THE EARTH, ACCORDING TO UMGROVE (1942)

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MANUSCRIPT RECEIVED BY THE SECRETARY OF THE SOCIETY, MAY 7, 1946

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 60, PP. 1771-1784, 6 FIGS., 2 PLS. NOVEMBER 1949

VALLEY WIDENING PROCESSES IN THE COLORADO ROCKIES

BY BENJAMIN A. TATOR

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ABSTRACT

Studies of local widening in canyons of the Colorado Front Range near Colorado Springs show detailed evidences of the processes involved.

The streams of the region are typically ephemeral, subject to repeated floods from torrential runoff. These conditions are believed to have been uniform for a long period of time.

The widening results in small, flat-floored basins usually associated with tributary drainage junctions. The basin floors are thinly alluviated except for numerous bedrock exposures along modern and old channels.

The processes involved are: (1) retreat of valley sides by weathering and local corrasion by streams and (2) shifting of channels by plugging the main channel at the basin head and by tributary deposition on the basin floor. Repeated shifting of the channel over a long period and flood scour to an average depth produces a flat bedrock floor.

These valley widening processes are believed to be significant in the development of prominent piedmont flats in the region.

INTRODUCTION

This paper is the first of a series resulting from studies carried on by the author in the Colorado Springs region during the summers of 1940, 1941, 1946, 1947, and 1948, and the winter of 1945–46. This series of papers is intended to suggest an origin for the broad, alluviated interstream flats which are so conspicuous along the eastern face of the Front Range in that region. It is believed that these piedmont flats are caused by processes active in the modern valley bottoms. This paper will describe these processes.

Mackin (1948) arrived at somewhat similar conclusions at the time of completion of research for the present paper. The streams described are in geologic circumstance quite similar to those discussed herein. Mackin, however, related his concepts to accepted stream terminology, whereas this writer has preferred to avoid such usage. This decision is based on a desire to emphasize actual processes of valley widening in the Col-

orado Rockies unbiased by the clouding influence of modern terminology. The present author, however, is in complete accord with Mackin's belief that correct approach to the understanding of stream activity requires a long term view. It is also agreed that the smooth profile theorized to exist in the graded stream is non-existent and that accurate analysis reveals a segmented profile, each segment of which is adjusted to its own particular set of controlling factors.

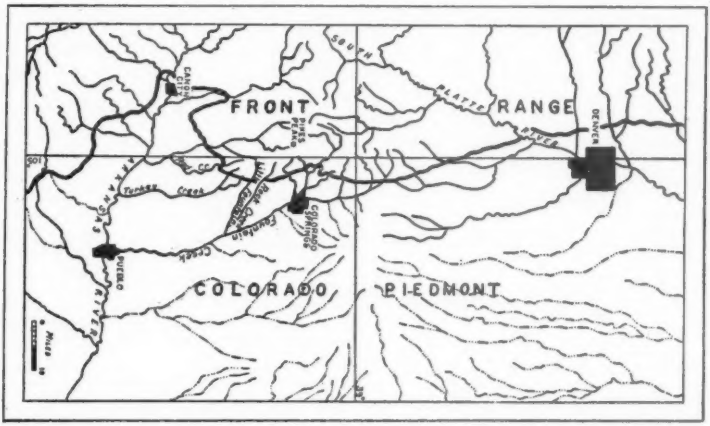

FIGURE 1.—DRAINAGE MAP OF COLORADO SPRINGS REGION AND ADJACENT PORTIONS OF THE FRONT RANGE

ACKNOWLEDGMENTS

The writer is indebted to Dr. C. J. Roy, former director of Louisiana State University's Field Camp and present head of the Geology Department of Iowa State College, for guidance in this problem. G. A. Stokes, C. L. Buttorff, and C. H. Lautier (assistants at the Field Camp) served as field assistants in various phases of the work. The author is also indebted to Dr. R. J. Russell, Assistant Director of the School of Geology, Louisiana State University, for helpful criticism and suggestions. Grateful thanks is extended to Mrs. R. W. Magee, Children's Librarian at the Colorado Springs Public Library, and to Mr. R. W. Magee, also of that city, for encouragement and material assistance.

GENERAL SETTING

Drainage

The majority of the valleys of the Front Range in east-central Colorado are steep-walled canyons developed in the bedrock of the mountain mass. Locally, however, small, thinly alluviated, flat-floored basins occur along these canyon segments. These basins are particularly well developed along the canyons of Rock and Red creeks, two tributary streams of the Arkansas River Drainage System (Fig. 1).

Red Creek rises in the Front Range and flows generally south to a junction with Beaver Creek, a tributary of the Arkansas River (Fig. 1). The segment considered here includes a small canyon basin located in the headwater portion about half of a mile upstream from the canyon mouth of this stream (Fig. 2). Red Creek rises at an elevation of approximately 8500 feet, dropping to an elevation of 7000 feet in the 2¼ miles to its canyon mouth at the mountain front. The average gradient is 600 feet per mile through this segment. The gradient in the canyon basin portion, however, is approximately 400 feet per mile.

Rock Creek originates in the Front Range and flows generally southeast to join Little Fountain Creek, a tributary to Fountain Creek, in the piedmont beyond the mountain base (Fig. 3). The stream rises at an elevation of approximately 8500 feet and the canyon mouth is at an elevation of 6500 feet, representing an average gradient through the canyon segment of 650 feet per mile. The gradient through the canyon-mouth portion is 400 feet per mile.

Climate

This region is semiarid, which results in rather specialized types of geologic activity along the modern drainage lines. Heavy localized precipitation and torrential runoff are characteristic.

The mean annual rainfall varies from 12 inches in the lower piedmont, 15–20 miles from the mountain base, to 14 inches in the upper piedmont and on the lower mountain slopes. A prominent feature of this precipitation is the concentration of the greater amount in the warmer part of the year, more than 75 per cent falling between April 1 and September 1. Meteorological data from the Colorado Springs Weather Station, collected over a 59-year period, substantiate this fact.

Average Monthly Precipitation in inches
(U. S. Weather Bureau)

January	0.22
February	0.36
March	0.72
April	1.57
May	2.21
June	1.82
July	2.94
August	2.29
September	1.08
October	0.64
November	0.35
December	0.31
	14.51 inches

An important precipitation characteristic is high rainfall intensity during local summer thundershowers which results in disastrous floods. The Colorado Springs Gazette (June 27, 1884) describes one such occurrence as follows:

"There had been several thunder showers and the creeks were somewhat swollen, . . . But about 4 o'clock . . . a heavy cloud came up over Cheyenne Mountain . . . sundown, . . . rain and hail began to fall in tremendous torrents . . ., the whole surface of the country was flooded as though it were a vast lake, and in some of the ravines the water rushed along in torrents 20 to 30 feet deep. The storm continued in full violence until about 9 o'clock. The area of the storm was confined within . . . a radius of 3 to 4 miles."

The same paper (August 1, 1885) reports a storm which flooded Monument Creek, a small headwater tributary of Fountain Creek, but effected the flow of Fountain Creek very little. Another storm, extending over an area from Colorado Springs southward to a point beyond the Town of Fountain on May 27, 1902, flooded Fountain Creek and temporarily deepened its

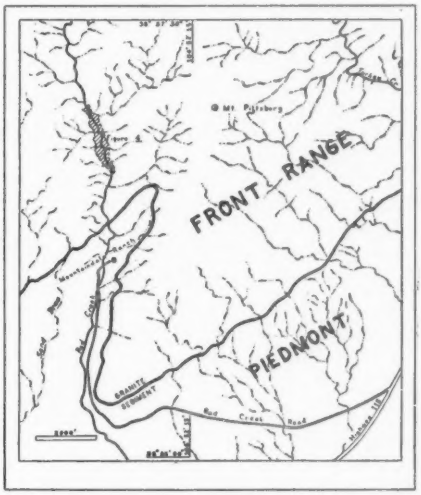

FIGURE 2.—DRAINAGE MAP SHOWING LOCATION OF THE CANYON-BASIN IN THE MOUNTAIN CANYON SEGMENT OF RED CREEK

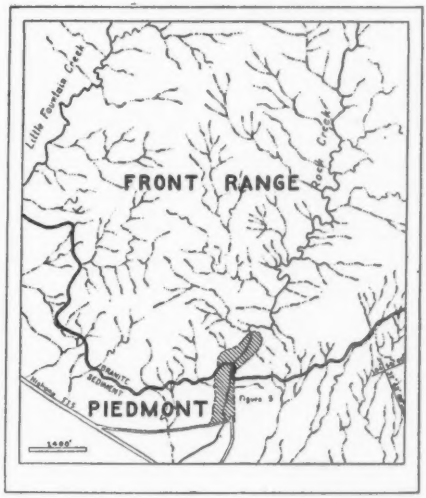

FIGURE 3.—DRAINAGE MAP SHOWING LOCATION OF CANYON-MOUTH PORTION OF ROCK CREEK AT THE LOWER END OF THE MOUNTAIN CANYON SEGMENT OF THAT STREAM.

channel 30 feet in places. The precipitation causing this flood, from a number of local thundershowers, totaled 3.02 inches.

The localized torrential runoff from these storms is a capable scouring agent in the channels affected. Numerous references to scour activity of varying magnitude are available in the literature. Unfortunately, however, no records have been kept of scouring action by floods in the valleys of Red and Rock creeks. Most of the references are concerned with flood scour in the valley of Fountain Creek. The flood of May 30, 1935 perhaps best typifies these occurrences. In a series of local thundershowers which occurred throughout the Fountain Creek Basin, ranging from 3 to 6 inches in magnitude, the various tributaries were swollen by floods. The channel of Fountain Creek was scoured to a depth of 3–9 feet below the channel alluvium at this time. A scouring of similar magnitude is recorded for the flood of June, 1921.

Characteristics of Foothill Canyon Segments

General Description

The drainage lines of the Front Range occupy deep, steep-walled ravines and canyons in the igneous and metamorphic rocks of the mountain mass. These rocks, Precambrian in age, have been intensely metamorphosed and exhibit a marked uniformity in their resistance to weathering and erosion; the dendritic character of the drainage net attests to this. The valley bottoms are floored with bedrock, covered in some portions by a relatively thin veneer of alluvial material. In several of the larger canyons, flat-floored, alluvium-covered basins produce local widenings in the otherwise narrow canyons. These basins are invariably located where numerous tributary streams join the main stream. The alluvium covering the bedrock of the basin floor is coarse-textured, most of the larger fragments being subangular to subrounded. A typical basin located in Red Creek Canyon (Fig. 2), was mapped in detail by the author (Fig. 4).

Red Creek Basin

The present stream channel occupies a variable position on the basin floor, in some places impinging against the canyon wall (Fig. 4, Ci; Pl. 1, fig. 1), in other places out in the basin (Fig. 4, Cb). The channel varies from 2 to 6 feet deep, here and there exposing bedrock. Additional dissection of the basin flat is provided by older channels, now abandoned, which also vary up to 6 feet deep. Some of these are fresh with steep, unmodified banks (Fig. 4, Au); others are shallower and present a more rounded appearance (Fig. 4, Am; Pl. 1, fig. 2). The channel pattern of the basin floor is interlacing; divergences occur both from the main channel (Fig. 4, Cd) and from other, older channels (Fig. 4, Ad).

The smoothness of the valley flat is interrupted by alluvial fans of varying size, rock trains, and boulder plugs. Fans have been deposited by both the main stream and tributary streams. Some mainstream fans have apices that extend into the constricted portion of the canyon at the upper end of the basin (Fig. 4, Fa).

Large parts of fans have been removed by dissection, leaving remnants in interchannel areas. Much fan dissection has been accomplished by the main channels of the valley flat, as is demonstrated by scars in the dissected fan escarpments and by the old channels along the bases of the fans. Some, however, is caused by tributary stream activity. The dissection exposes in places the underlying bedrock of the basin floor. Fan sequences occur below the mouths of larger tributary gullies (Fig. 4, Fo, Fy). The highest of these sequence remnants present steep 10- to 15-foot alluvial escarpments toward the basin center (Pl. 2, fig. 1). It is often possible to reconstruct the events which resulted in development and dissection of the local fan sequences.

The rock trains are elongate ridges of coarse bouldery material, usually aligned along the interlacing channels of the basin floor (Fig. 4). These ridges, somewhat similar in appearance to natural levees, project from a few inches to several feet above the alluvial flat (Fig. 6).

The basin walls are steeply sloping exposures of mountain complex bedrock. They exhibit a marked constancy in angle of slope. The bedrock surfaces are covered by only small quantities of angular rock fragments in transit. A sharp angle ordinarily marks the break between sidewall and basin floor (Pl. 1, fig. 1). In

FIGURE 1. MODERN CHANNEL IMPINGING AGAINST THE EAST WALL OF RED CREEK BASIN
This view is north, as shown by the arrow in Figure 4. The basin wall is in the right of the view and the valley flat extends to the left.

FIGURE 2. ABANDONED CHANNEL IN THE VALLEY FLAT OF RED CREEK BASIN
The view is north, as indicated by the arrow in Figure 4. The channel has been somewhat modified by surface processes active since abandonment. Note the flat aspect of the basin floor.

FIGURE 1. FAN SEQUENCE ALONG THE WEST SIDE OF RED CREEK BASIN
At the mouth of a large tributary canyon. Trees are observable on the dissected fan escarpment appearing in the middle of the view. Tributary channels have been cut in this old fan, as shown at both right and left. Each now contains a younger fan. Note the old stream scars at the base of the escarpment in the center. This view is west from the basin center.

FIGURE 2. VIEW SOUTH FROM MODERN CHANNEL BOTTOM IN THE NORTH END OF RED CREEK BASIN
The east valley wall is in the left of the view. The boulder plug (A of Figure 6) is in the middle, old channel segment (1) extending downvalley just beyond this obstacle. Size relationship may be gained by comparison with the 12-inch pine tree just left of the middle of the view.

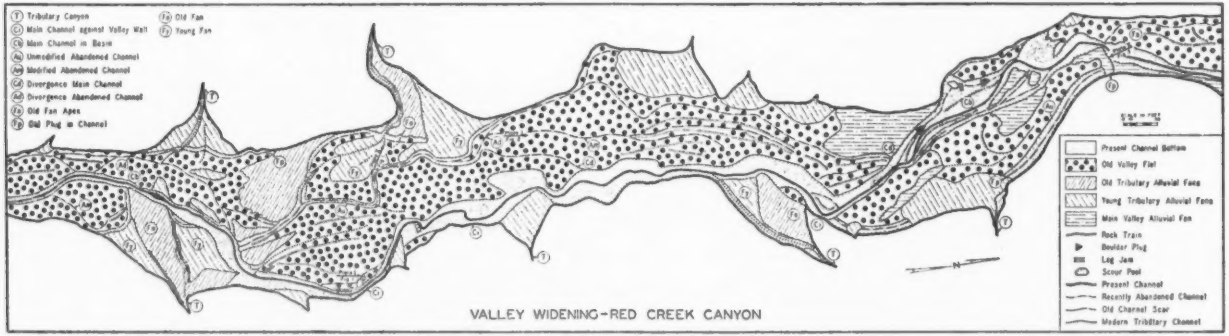

FIGURE 4.—DETAIL MAP OF RED CREEK BASIN

places, the contrast is emphasized by the presence of channels or channel scars. Talus and fan accumulations, however, obscure the angle between basin wall and basin floor at many places.

Characteristics of Canyon-Mouth Segments

General Description

The valley bottoms at the mountain front are flat and veneered with a blanket of coarse fanglomerate up to 12 feet thick. The present channels occupy variable positions with respect to the valley sides. In general these valleys flare outward from the mountain base in the down-valley direction, though local narrowings and widenings occur in many. The interstream areas in the piedmont at the mountain base consist of thinly alluviated bedrock surfaces herein termed piedmont interstream flats. The bedrock of the mountain base portion of the piedmont consists of homoclinal sedimentary strata dipping in general toward the east. Many of these piedmont valleys contain steplike flights of thinly alluviated, rock-floored terraces along their walls. Where these occur they border the interstream flats, forming "steps" in the scarped edges of the latter features.

The mountain front portion of Rock Creek valley was mapped in detail by the author (Fig. 5).

Canyon-Mouth Segment of Rock Creek Valley

The modern channel is variable in position and is incised from 2 to 8 feet below the alluvial surface (Fig. 5, Ci, Cv). Numerous unoccupied channels which form an interlacing system on the valley flat (Fig. 5, Cd, Ad) compare in every detail with those in the Red Creek basin. Many show evidence of recent occupancy (Fig. 5, Au); others have been idle long enough to display considerable modification (Fig. 5, Am). Bedrock is exposed in a number of places in these channels, in all portions of the valley and at approximately the same depth below the alluvial surface. This bedrock surface appears to be relatively flat.

Rock trains, boulder plugs, and alluvial fans occur in the valley flat with relationships and characteristics similar to those already described. Because of the more open character of the canyon-mouth segment of Rock Creek valley, however, the fans are larger (Fig. 5, T), and the boulder trains more continuous than in the mountain canyon basins. Tributary fans form plugs across abandoned channels of the valley bottom in several places (Fig. 5, Fp).

The curved arrow symbol in Figure 5 indicates the approximate position of the sediment-granite contact where it crosses the valley of Rock Creek. This is also the line of contact between mountain mass and piedmont. The contact locally follows a thrust fault, the west side granite of the upthrown block and the east side sediments of the downthrown block, in this instance Upper Cretaceous, Pierre shale. The granite walls upstream drop precipitously to the valley flat, meeting it at a sharp angle in most localities. Farther down valley, in the sedimentary section east of the fault, the sidewalls are steep but lower. Fanglomerates of the piedmont interstream flats cap the valley walls, whose lower portions consist of the soft, easily weathered Pierre shale.

One terrace level is present in this segment of Rock Creek valley (Fig. 5), about 5 feet above the modern valley flat, merging with the flat near the mouth of Rock Creek Canyon in the upstream direction. Above the terrace, mountain base portions of piedmont interstream flats bound the valley on each side; their north and south sides standing respectively at 50 and 15 feet above the valley floor. The terrace flat is similar to all respects to the modern valley flat.

Mechanics of Valley Widening

Climatic Factors

In a semiarid region where precipitation yields intense, localized runoff of torrential nature usually confined to a limited portion of one drainage line, large quantities of rock debris are shifted for relatively short distances. Though intensity of the runoff differs, a certain mean intensity prevails over long periods of time. The rate of weathering on exposed surfaces of homogeneous rock is rather constant, if means for removing the debris from the exposure exist. The amount of debris supplied by

FIGURE 5.—DETAIL MAP OF CANYON-MOUTH SEGMENT OF ROCK CREEK VALLEY

weathering remains comparatively uniform with long-continued uniformity of climatic conditions. Hence, weathering and erosional activity in this region are somewhat uniform.

Retreat of Slopes

A number of students of arid and semiarid land-forms agree that slopes weather back at a constant rate and at a relatively constant angle of slope where similar conditions of climate, lithologic character, and rock structure prevail. Among these are Lawson (1915), Bryan (1936), Rich (1935), Field (1935), and Gilluly (1937).

Bryan (1936) noted that weathering controlled by widely spaced joints yields steep slopes in massive rocks, whereas more closely spaced systems of joints yield lower slope angles. Possibly localized joint systems in the granitic rocks of the Front Range are responsible in part for the location of the canyon basins along the drainage lines of the region.

Field (1935) noted that many factors control the retreat of slopes, including the length of the slope, the size and shape of the weathered rock fragments, and their rate of loosening. He concluded that rill wash on the rock surface and gravity are the important factors in the removal of material from the base of the slope.

Gilluly (1937) agrees with Bryan and Lawson (1915) that there is a correlation between the average size of the rock fragments weathering from the slope and the angle of slope.

Rich (1935) observed that slopes in semiarid climates will retreat at a constant angle so long as the weathered material being supplied from higher up on the slope is removed from the base by running water. Accumulation of debris at the base, however, will cause at least a temporary halt in the escarpment retreat locally.

The sidewalls of the stream valleys in the Colorado Springs region exhibit a marked constancy of slope angle. The rate of retreat, however, varies from place to place. The more rapid retreat of canyon-basin walls, as compared with the walls of the narrower canyon segments, may be related to a more favorable arrangement of secondary rock structures. The greater number of tributary streams which enter the basin segments are a perhaps related factor.

A more striking contrast is the retreat of slopes in the sedimentary rocks of the piedmont as compared with slopes in the granitic rocks of the foothills. The piedmont valleys are much wider than their canyon portions upstream. Variations in valley width in the piedmont are related in every case to the bedrock forming the valley sides; valley segments are, in general, much narrower in the more resistant sediments than in the less resistant ones.

Valley widening is accomplished in this region by the following processes: Spalls weathered from the valley walls, moved by water and gravity, collect at the base of the slopes. Periodic flushing, by torrential runoff, along channels which impinge against the base of the sidewalls keeps these portions free of debris accumulations which would prevent uniform weathering and sidewall retreat. Tributary streams, particularly in the more easily eroded sediments of the piedmont, are an additional factor in sidewall retreat. These work headward into the sidewall slopes, wearing them back and moving the eroded material onto the valley flat where it may be removed by the main-channel activity.

Channel Shifting

The abandoned channels characteristic of the valley flat mark positions of the main channel at various times during the valley widening (Figs. 4, 5 Au and Am). These shifts in channel position are the result of alluviation on the valley floor during torrential floods. Such aggradational activity includes the development of rock trains, boulder plugs, and associated log jams, and alluvial fans.

Tolman (1905–1906) used the terms "arroyo train" and "rock train" to describe extensive radial deposits of bouldery material on the bajada surfaces of the Southwest, concluding that they are stream deposits. Matthes (1930), observing similar features in the Yosemite region, stated that these are not natural levees formed during times of overflow but material thrown bodily out of the torrent channel. He described fragments in the rock trains much larger than any observed by this writer in the Rocky Mountain foothills region; this would not appear to be a diagnostic difference, however. It is this writer's opinion that, in the Colorado Springs region of the Front Range and adjacent pied-

mont, these are torrent levees representing an accumulation of bed-load fragments which managed to escape the main thread of flood current during torrential runoff. The term "rock train" is deemed appropriate, and is inoffensive because it carries no implication of origin.

The rock trains act as confining agents for runoff, but they also prevent new courses of floodwater from re-entering channel segments once abandoned. Breaching or overtopping of a rock train during an excessive flood allows the runoff to escape from older established channels into new paths across the valley flat. Segments of channels are thus abandoned and the upper ends of the abandoned portions are then alluviated. Slack water conditions after the main flow has proceeded down a new course favor this alluviation. Alluvial plugs toward the heads and rock trains along the sides of the abandoned channel segments render it unlikely that a channel segment, once abandoned, will be reoccupied. Diverted runoff assumes courses determined largely by the rock train pattern of the valley flat. Shifts in channel position to the interchannel areas between rock trains or between rock trains and adjacent valley walls are more probable. Because of this shifting, flood runoff is not likely to follow the same course for any appreciable number of years. Diversions shift the channels so effectively as to subject all portions of the valley flat to channeling at some time or other.

Boulder plugs and associated log jams are important agents of channel shifting, as has been noted in Arroyo Seco Canyon, Los Angeles County, California, by Krumbein (1942). In an earlier paper (1940, p. 645), Krumbein describes channel shifting due to the presence of plugs and jams.

> "The upbuilding of the beds forced the stream to cut laterally around the jam unless the jam itself was partially undermined by the clear water pouring over its edge. In either case a channel lower than the surface of the deposit was soon developed."

Flood deposits similar to those described by Krumbein are numerous along the valley bottoms of the foothills and the piedmont of the Colorado Springs region. Abundant diversions in the valleys of Rock and Red creeks are directly attributable to boulder plugs and log jams; the two are often associated features (Figs. 4, 5).

The main valley fans promote channel shifting. Because surface runoff can assume any radial down-slope direction from the fan apex, flood channels commonly impinge against either valley wall below or marginal to the fan. Fan deposits often cover older channels, thus masking the complexities of the channel patterns. An actively growing fan tends to shunt earlier drainage aside into new channels. A valley fan of this nature is represented by dissected remnants at the north end of Red Creek Basin (Fig. 4).

Tributary fans force the main channels toward the opposite side of the valley from their site of deposition. Their development favors slack water conditions upstream along the main channel, leading to alluviation of the main valley flat. Blocked channels may force the runoff to cut back into alluvial material and talus along the valley wall at elevations well above the former channel bottom. A number of the stream-cut escarpments in the fan sequences along the valley sides are probably developed in this manner (Pl. 2, Fig. 1).

Diversion of the main channel by the deposition of boulder plugs and log jams, as well as valley fans, is well exemplified in the north end of Red Creek Basin (Fig. 6) where at least five diversions are apparent. The oldest channel segments, designated (0) in the figure, are greatly modified to shallow, elongate scars in the alluvium of the modern valley flat. The other channels are much fresher, showing evidence of quite recent abandonment. The oldest of these, (1) in the figure, splits off from the modern channel in the constricted north end portion of the basin. Its diversion and abandonment are apparently associated with boulder plug deposition. The upstream end of channel 1 is now completely blocked by such a boulder plug (A). During a subsequent period of dissection, the runoff developed channel (2). Later channel plugging action at (B) forced the runoff into the short segment (3). A plug then developed at (C) leading to the occupancy of channel (4) and the abandonment of segments (3) and (2); the runoff at this point had been shifted to the opposite side of the valley from the original position along channel (1). Later

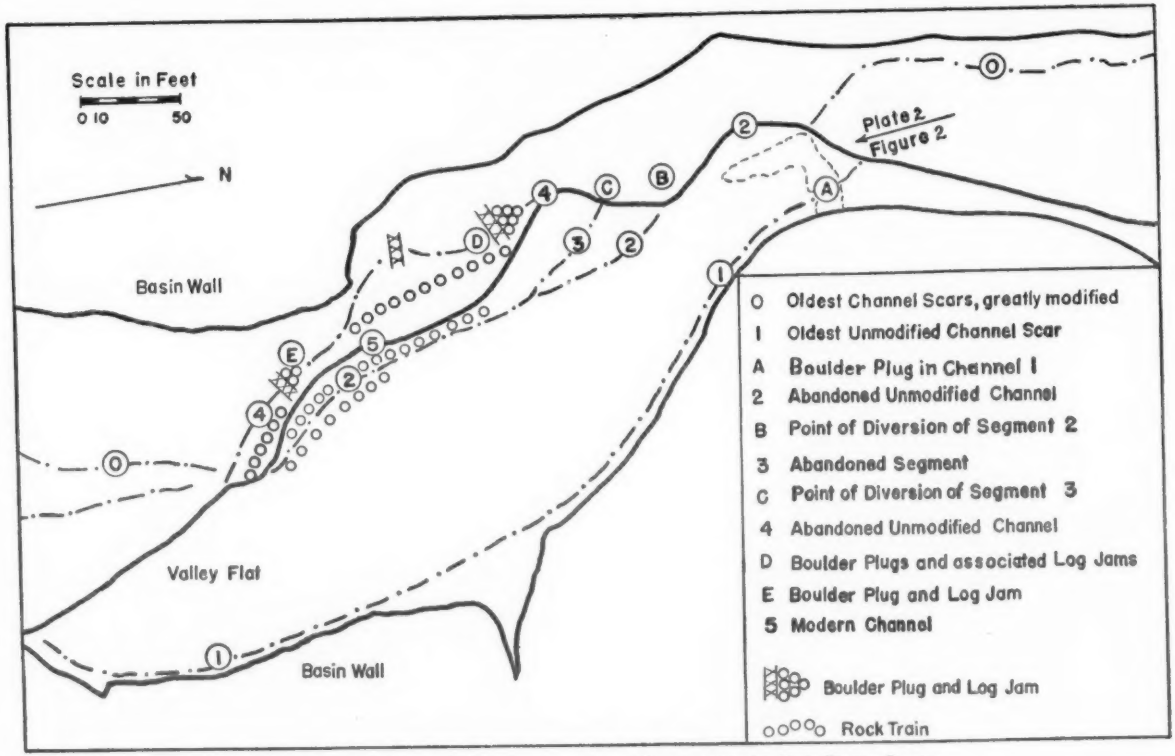

FIGURE 6.—MAP OF DRAINAGE DIVERSIONS IN THE NORTH END OF RED CREEK BASIN

log jam and boulder plug development, (D) and (E), has shifted the course to segment (5), the modern channel.

Krumbein (1942, p. 1400) notes that flood deposits are caused by several factors including:

> "abrupt decrease in gradient as at tributary mouths; abrupt velocity decrease associated with locking logs and boulders; velocity decrease caused by water profile adjustments at constrictions; and perhaps abrupt velocity decrease caused by hydraulic jumps."

These factors may operate singly or in combination at any time during the flood, causing local alluviation and channel shifting. Velocity decrease, when flood waters move into broader portions of the valley, occurs in these canyon-basin segments. The valley fan represented by dissected remnants in the north end of Red Creek Basin (Fig. 4) is a floodwater deposit made by the runoff as it "spread-out" from the narrower canyon segment upstream. A contributing factor to such alluviation is the network of old channels which dissipate the energy of the floodwaters by dividing them, thereby reducing their transportational ability and causing deposition. Thus, as the flood moves downvalley, the creation of alluvial obstacles shifts the locus of channel activity about on the valley floor. This shifting plays an important role in the process of valley widening.

Bedrock Erosion

Floods are effective agents of erosion. Their waters pick up the loose debris and scour channel sides and bottoms even after exposing the solid bedrock of the valley floor. Kesseli (1941) points out that even a stream loaded to capacity will continue to attack the stream bed by the impact of material carried in the bed load. Thus, even when the flood is heavily charged with debris, vertical corrasion will be favored. Kesseli notes further than during the last stages of flood the torrent changes to a "brook" and, despite decreased volume, cuts a trench into the flood deposits. Such cutting, if extensive enough to scour through the flood deposit, will reach the underlying rock surface. Shifting channels that are capable of cutting as deep as the bedrock are capable of scouring any part of the valley floor, eventually tending to corrade the bedrock surface to approximate flatness.

The lack of accounts of local flood scour in the tributary valleys of this region make it impossible at present to arrive at an average figure for the effective depth of such action. Bailey (1934), however, in studies of flood activity in northern Utah, a region climatically similar to the Colorado Springs region, offers the following data. Two floods in Ford Canyon in 1923 and 1930 cut a channel 40 feet deep in the fan of that creek at the canyon mouth. Parrish Creek in the same area, during the 1930 flood, cut down 70 feet in the canyon fill and exposed stretches of bedrock in the channel bottom.

Woolley (1946), in detailed studies of flood activity in the Wasatch Range, Utah, found that Kanab Canyon was visited by torrential floods each summer from 1882 to 1886. In that period, the channel depth was increased by 50 feet in places.

Frankenfield (1923, p. 421) observed of flood activity in Willard Canyon, Utah, that: "The stream channel was scoured to bedrock . . . , the exposed surfaces showing much watermarking, as if by previous floods, . . ."

These figures show that, in other regions affected by torrential runoff, scour activity reaches to considerable depth in the valley alluvium. Furthermore, in none of the valley bottoms of this region is the alluvium more than 25 feet thick; and in most instances, it is less than 15 feet. The numerous old stream channels, many of which expose bedrock in their bottoms at scattered localities, suggest that the scouring ability of torrential runoff affects the bedrock widely. Additionally, none of these channels is scoured more than a few feet into the underlying bedrock at any locality in the valley flat.

Bryan (1940) described a process of channel shifting and scouring which he called "gully gravure". This process requires porous debris which becomes impervious as it weathers, thereby increasing surface runoff. This material is gullied during torrential runoff which, if of great enough intensity, may scour to bedrock. The new gully then fills with fresh alluvium which is pervious and inhibits runoff, and subsequent gullying action shifts to the older, more impervious debris adjacent. Thus, gullying shifts laterally, and parallel incision to bedrock to about the same depth may occur each time, if

the alluvial debris is not too thick for scour to reach the underlying bedrock. Bryan applies this process to the retreat of slopes, if there is a supply of rock fragments from higher areas which will be carried by storm runoff over a slope developed on material which weathers into a clayey mass. These special conditions are met in many of the stream valleys in the piedmont portion of the Colorado Springs region. For example, the piedmont segment of Rock Creek valley is developed on a soft, fissile shale (Pierre shale) which weathers to a clayey mass, and there is a continual source of rock fragments in the fanglomerates of the piedmont interstream flats which form the upper portions of the valley walls. This may be one manner by which sidewalls retreat locally in the piedmont, and account in part for the greater width of these valley segments.

Conclusions

The streams of the Colorado Springs region are actively degrading their valleys. This conclusion is supported by the deep incision of the drainage lines in narrow canyons in the bedrock of the mountain mass; the steep gradients of the streams, 300–1300 feet per mile in the mountain portions; and the bedrock exposures along the valley sides and in the broad flats of the piedmont areas. Furthermore, these streams do not exhibit smooth profiles, being segmented by numerous changes in gradient with falls and rapids along their extent. In a degrading region such as this, the alluvium of the valley bottoms is continuously removed at a rate about equal to its replenishment by weathering and erosion. In order that degradation be continuous, the thickness of valley bottom alluvium must be slightly less than the vertical distance that the average flood is capable of scouring. If the average depth of scour is somewhat less than the depth of alluvium, the underlying bedrock is not within the scope of stream erosion. On the other hand, if the average depth of scour greatly exceeds the alluvial thickness, the excess energy of flood runoff will keep the valley bottom swept clean of alluvial debris, and the flood channels would be deeply incised in the bedrock floor. Neither of these conditions obtain. The channel depths in the majority of these valley flats reveal bedrock at many localities in their extent, but the channels are not incised to great depth below the bedrock surface.

Allowing then that the bedrock of the valley floor lies just within the scouring scope of the average flood, there is a means of producing a relatively flat bedrock floor in segments of these valleys where conditions promote sidewall retreat. Continued shifting of the locus of scour over the valley bottom effectively corrades the underlying bedrock at a depth about equal to the alluvial thickness. Though an average depth of scour will prevail under uniform climatic conditions over a long period of time, individual floods will vary somewhat in intensity. These variations should be indicated by minor protuberances above and minor channel irregularities below the bedrock surface. The numerous bedrock exposures in the irregularly-trending channels of the valley flats show these indications. In places the bedrock is incised a few feet below its upper surface; in other localities the bedrock lies several feet below the channel-bottom alluvium. The field facts favor the presence of minor irregularities in the bedrock floor of the valley flat.

Channel shifting, by which the valley walls are periodically freed of the cloaking effect of debris accumulations along their base, is also a means of local lateral corrasion, whenever the active flood channel impinges against the bedrock of the sidewalls.

Valley deepening and valley widening are occurring simultaneously, as shown by locally smooth stream profile which extends without major irregularity through both narrow canyon and adjacent broader canyon-basin segments. Locally, however, the retreat of valley walls is the more rapid process, so that canyon-basins and broad piedmont valley flats are being developed.

In the writer's opinion, the valley bottoms of the region are being lowered uniformly but slowly, while local widening is occurring by channel shifting and sidewall retreat. The downcutting is not accomplished by continuous scour, but rather by the lateral shifting of the locus of corrasion by sporadic flood runoff. This results in the uniform lowering of the bedrock floor in all portions of the valley bottom. Deep incision along any particular thread of flow cannot occur unless some means of permanently confining the channel is developed.

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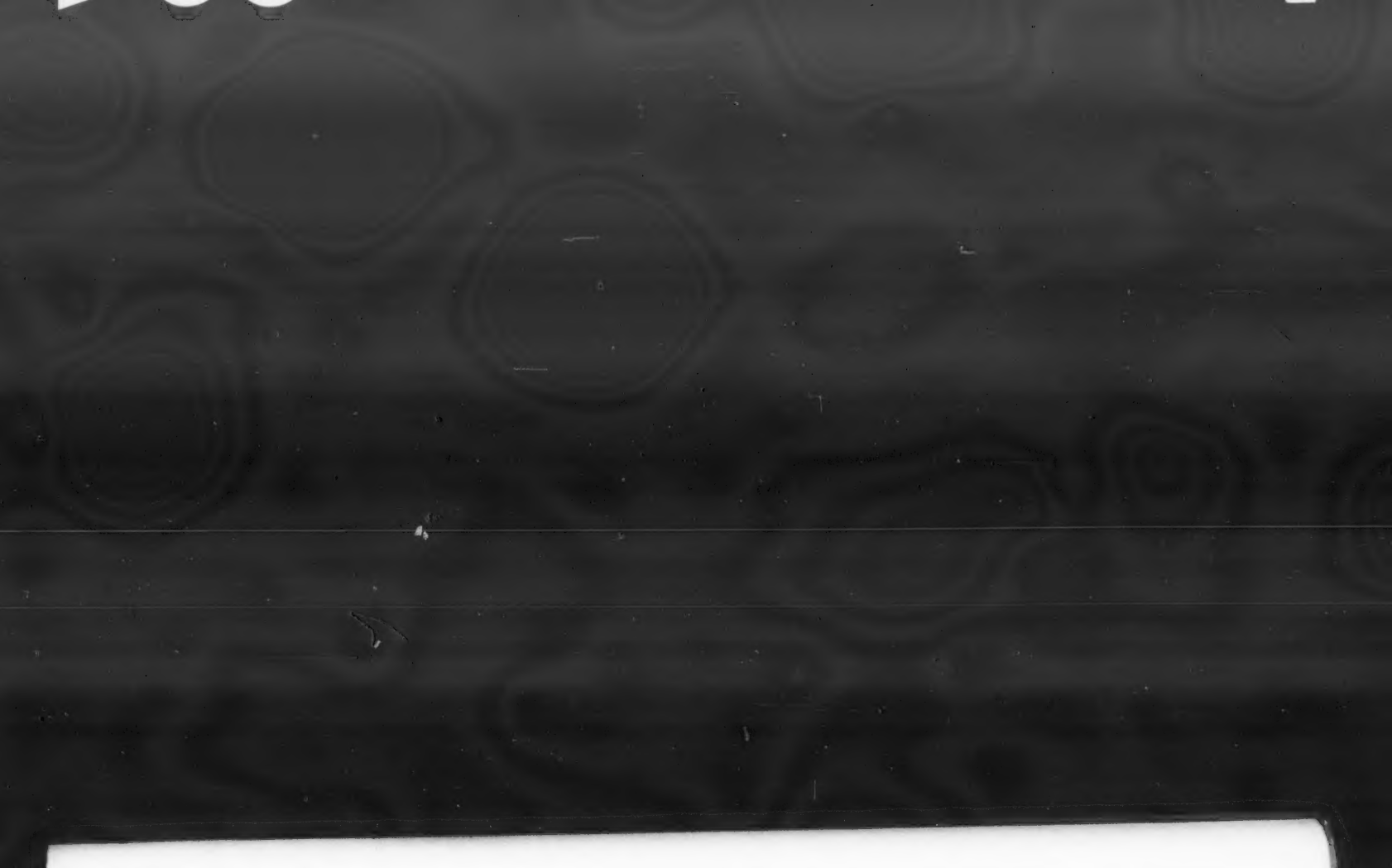
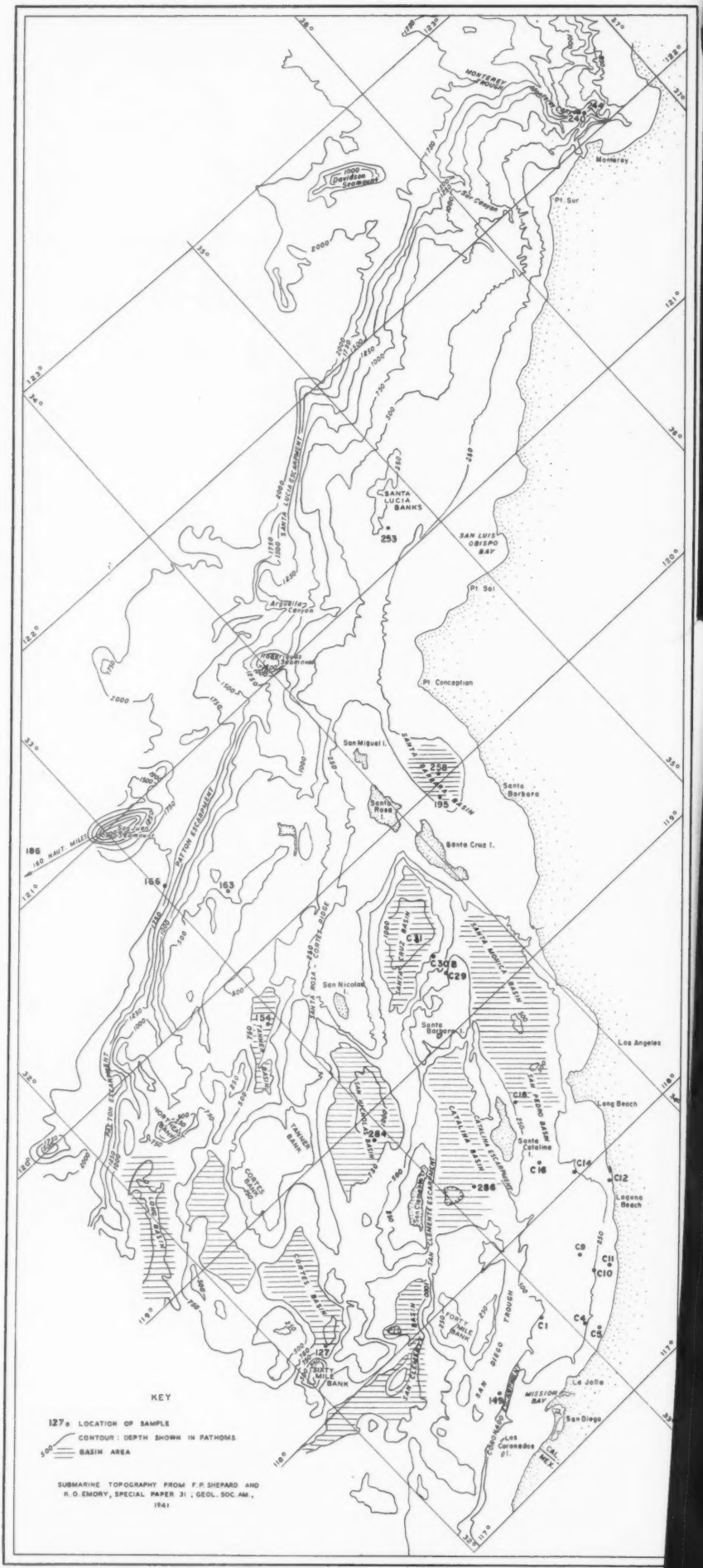
Louisiana State University, University, Louisiana

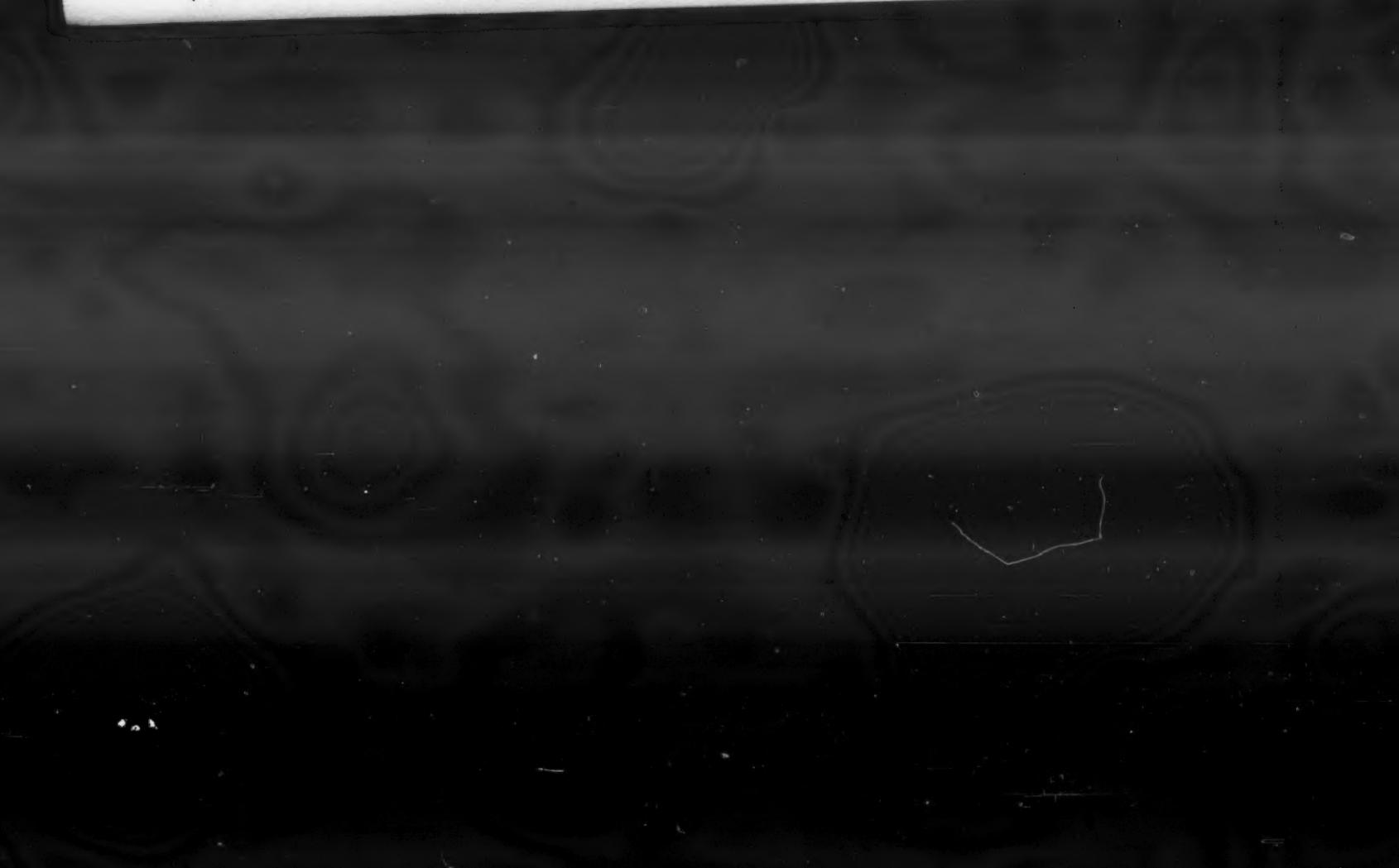
Manuscript Received by the Secretary of the Society, November 13, 1948.

Louisiana State University Geology Camp Research Paper No. 5





LOCATION OF SAMPLES FROM THE FLOOR OF THE PACIFIC OCEAN OFF THE COAST OF CALIFORNIA



BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 60, PP. 1785-1808, 22 FIGS., 5 PLS. NOVEMBER 1949

CLAY MINERAL COMPOSITION OF SOME SEDIMENTS FROM THE PACIFIC OCEAN OFF THE CALIFORNIA COAST AND THE GULF OF CALIFORNIA

By Ralph E. Grim, Robert S. Dietz, and W. F. Bradley

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Abstract

This report presents the results of a study of the clay mineral composition of a series of bottom core samples collected by the staff of the Scripps Institute of Oceanography of the University of California in Pacific Ocean off the California coast and in the Gulf of California. The clay minerals were studied by X-ray diffraction, differential thermal, optical, chemical, and electron microscopic methods following particle size fractionation by sedimentation and supercentrifuge procedures.

All of the samples from the Pacific Ocean and from the Gulf of California contained illite, montmorillonite, and kaolinite; generally illite was most abundant and kaolinite least abundant. A chloritic clay mineral could be identified definitely in some samples. The clay minerals were in very complex mixtures, including mixed crystallizations as well as mechanical mixtures of discrete phases. In general the crystallinity was lower, the individual size smaller, and the intergrowth more intimate than in ancient sediments which have been studied by the authors.

Small amounts of quartz were associated with the clay minerals in the 1 to 0.1 and minus 0.1 micron fractions. Small amounts of another nonclay mineral crystalline phase that is probably a feldspar were also found in the finest size fractions of many samples.

The analytical data suggest that kaolinite is slowly lost during diagenesis under marine conditions, perhaps being changed to an illite or chloritic clay mineral. The data also afford evidence that potash is taken up by the clay and suggest that it is taken up largely by partially degraded illite which is carried into the sea. Magnesium also appears to be taken up by the clay, perhaps by the illite.

The widespread occurrence of montmorillonite indicates that this clay mineral is not lost quickly if at all during diagenesis under marine conditions.

Introduction

General Statement

This report presents the results of a study of the clay mineral composition of a series of core samples collected by the staff of the Scripps Institute of Oceanography of the University of California in the Pacific Ocean off the California coast and in the Gulf of California.

Clay mineral investigations in recent years have provided considerable data on the occurrence of clay minerals in soils and in ancient sediments, on the structure and properties of clay minerals, and on some of the conditions controlling their formation and their variation in properties. Information on present-day sediments is exceedingly scant but is essential to a complete understanding of the occurrence, origin, history, and properties of clays. Information filling in the gap between our knowledge of source materials (soils) and ancient sediments should provide keys for the paleogeographic interpretation of ancient sediments which would be of great value not only scientifically but also commercially, as for example by locating probable source beds of petroleum.

In this investigation no attempt was made to study any characteristics or components of the samples other than the clay minerals. Revelle and Shepard (1939) have published an excellent statement of the general character and distribution of the sediments off the California coast. Various special features of the sediments of the area have been studied by Trask (1931), Galliher (1932), Cohee (1937), Shepard and Wrath (1937), and Shepard and MacDonald (1938).

The submarine topography of the Pacific area has been considered in detail by Shepard and Emery (1941) and of the Gulf of California by Shepard (in press). Other data concerning the general oceanographic conditions of the areas are given by Sverdrup and Fleming (1941).

Acknowledgments

This work was financed by a grant from the Penrose fund of The Geological Society of America (268–39) for which appreciation is gratefully expressed. The staff of the Scripps Institute of Oceanography of the University of California, particularly H. U. Sverdrup, Director, Roger Revelle, and F. P. Shepard, willingly made available their samples and supplied any available information that was desired concerning them. Thanks are also due to these individuals for reading and commenting upon the manuscript.

W. A. White and L. D. McVicker, both of the Illinois Geological Survey, were the analysts for the differential thermal data and ultimate chemical data, respectively. The authors are deeply grateful to these men for carefully and

conscientiously carrying out jobs that were frequently laborious.

Finally the authors are indebted to M. M. Leighton, Chief of the Illinois Geological Survey, who generously made the Survey's facilities available whenever they were needed in the work.

Review of Literature

On the basis of his *Challenger* studies, Murray (Murray and Renard, 1891) suggested that the fine clay fraction of sediments on the present ocean floors was noncrystalline. At that time the notion that many clays were composed of amorphous components was widely held (Grim, 1942), largely because there were no adequate techniques to study such materials. Beginning about 1925, X-ray diffraction methods, improved microscopic techniques, and differential thermal methods were devised that permitted, for the first time, the determination of the crystallinity or noncrystallinity of the finest-grained fraction of clays and made possible specific identifications of the crystalline components.

Using the new methods, Revelle (1936) and Correns (Correns *et al.*, 1937) first showed that the fine-grained clay fraction of Recent marine sediments is substantially all crystalline. A start has been made identifying the clay minerals in the present ocean sediments but as yet the results are exceedingly scant, and some of the identifications are incomplete and questionable. The complete identification of the clay minerals in such materials is very difficult because they are composed of complex intergrowths of exceedingly small units of many clay minerals. Only in the last few years has the knowledge of clay mineralogy and the development of X-ray diffraction and differential thermal techniques permitted satisfactory identifications of the clay mineral components. For example the recent method (Bradley, 1945) of pretreating such clay material with glycol permits the identification of small amounts of montmorillonite by X-ray analysis that would have previously gone undetected. Even now, X-ray, optical, thermal, and chemical techniques must all be used together on such complex clay material before the mineral identifications can be trusted.

Correns and his colleagues (1937) found that mica (probably illite) was present in all of the bottom samples collected by the *Meteor* in the South Atlantic. In addition to the mica, the clay fraction of some of the samples was composed predominantly of kaolinite, in others of halloysite, and in a few of montmorillonite. A map was presented showing the distribution of the predominant clay minerals. Montmorillonite was limited to relatively small regions where there were neighboring areas of basic volcanic activity. Correns pointed out that the differentiation of halloysite and kaolinite is not certain. It seems probable from his data, that the sediments in the South Atlantic studied are chiefly mixtures of illite and kaolinite-type clay minerals. Montmorillonite appears to be limited to certain areas but this may be questioned because of the great difficulty of detecting small amounts of this mineral by the techniques available when Correns did his work.

Some of the *Meteor* samples were cores, and Correns (1938) reported no variation in any of his cores from the top downward. He found no evidence of diagenetic changes in the clay mineral components of his samples.

Dietz (1941) determined the clay mineral composition in 39 samples from the various oceans of the world. Most of his samples were collected by the *Challenger*. Dietz found illite in all of his samples; it was the predominant clay mineral in 23 of them. Illite was particularly abundant in the deep-sea clays. Kaolinite, although less abundant than illite, was present in all of the samples and predominated in 9 of them, most of which were relatively close to shore. Small amounts of montmorillonite were found in the near-shore samples, but it was not detected in most deep-sea sediments. Again it should be stated that adequate methods for detecting small amounts of montmorillonite had not been developed when Dietz did his work.

Dietz (1941) presented strong evidence for the development of illite on the sea floor in the Recent sediments. It is his concept that the illite develops largely by the alteration of montmorillonite in the source material carried to the sea.

Revelle's (1944) study of the Carnegie samples showed the finest clay fraction to be crystalline, but unfortunately his analytical

data did not permit a positive identification of the predominant clay minerals. In a few samples kaolinite or halloysite was definitely identified, but most of the samples are said to be composed largely of a "beidellite-like" clay mineral. Much of the material classed as beidellite earlier is now known to be a mixture of clay minerals (Grim and Rowland, 1942), and from Revelle's data it appears certain that most of his samples are mixtures of clay minerals, and that in many, if not most of them, illite is the predominant clay mineral.

Bramlette and Bradley (1940) made no serious attempt to determine the clay mineral components in the samples collected by the *Lord Kelvin* in the North Atlantic. They pointed out that the clay fraction possessed optical properties similar to those of beidellite or hydrous mica, and from their data it appears probable that their clay samples are made up of mixtures of clay minerals and that illite is an important component in many of them.

Kuenen (1943) obtained X-ray diffraction data for the clay fractions of a few of the samples collected by the *Snellius* in the Netherland East Indies area. He found muscovite (probably illite), kaolinite, and montmorillonite in each of the samples investigated. The relative amounts of the clay minerals varied in the different samples, but too few samples were studied to permit any correlation of clay mineral components with type of sediment or oceanographic conditions.

Collection, Location, and Description of Samples

The samples studied were cores collected during cruises of the *E. W. Scripps* using the gravity coring apparatus designed by Emery and Dietz (1941). (Pls. 1, 2; tables 1, 2, and 3).

The cores varied in length up to a maximum of about 9 feet. In general separate samples were studied from various depths in a given core in order to determine possible vertical variations in the clay mineral composition of the sea floor sediments. As Emery and Dietz (1941) and Piggott (1941) have pointed out, because of compaction and other effects, the thickness of the core is considerably less than the depth of sediment penetrated.

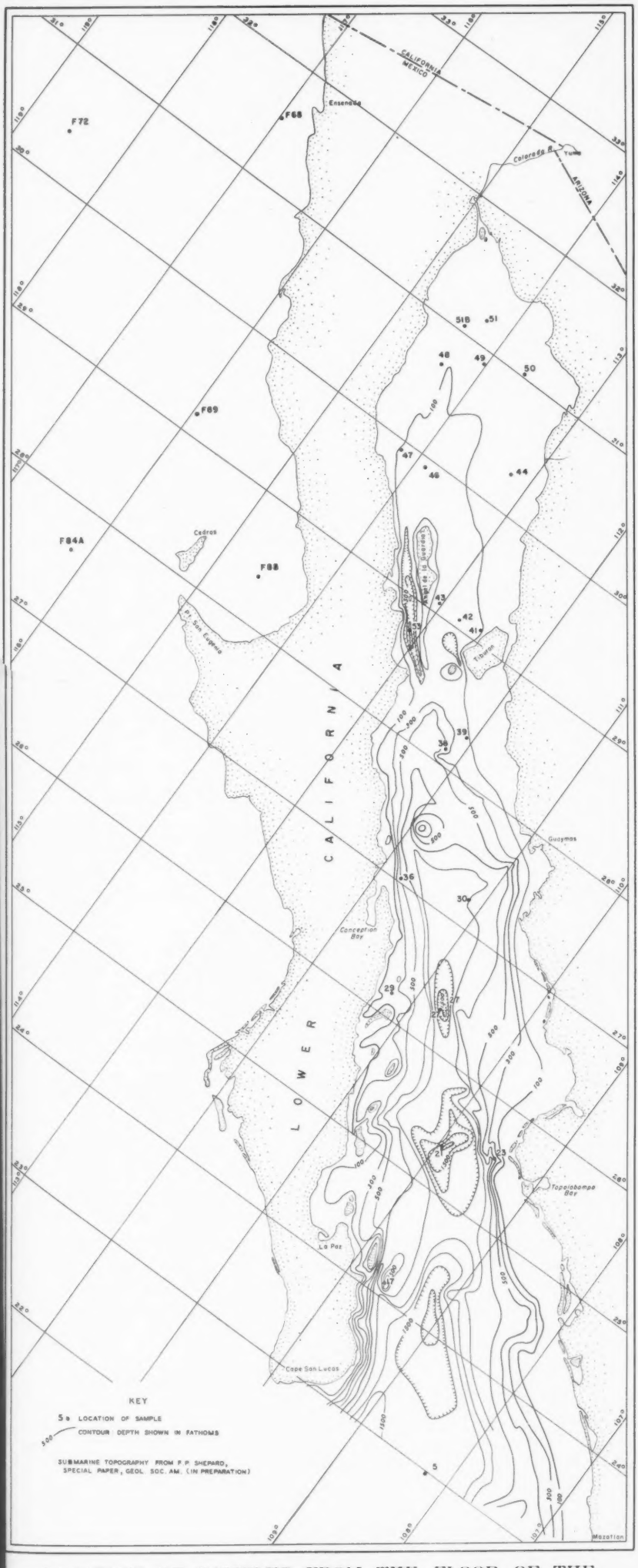
Analytical Procedure

Samples as received were undried and sealed in glass jars. The samples were first treated with 6 per cent H_2O_2 for 4 to 6 days to remove the organic material. If the samples contained carbonate they were then treated with 0.1N HCl until all effervescence ceased. The presence of appreciable organic material or carbonate in a clay frequently makes it impossible to obtain adequate diagnostic criteria to identify the clay minerals. Considerable work in the Illinois Geological Survey laboratory and by soil investigators has shown that H_2O_2 will oxidize the organic material in many soils and that it has substantially no effect on the silicate minerals. The acid leach was too mild to have an appreciable effect on the clay minerals in these sediments. In order to check this point, determinations were made of the amount of aluminum and silicon in the HCl leachate. In no case was a significant amount of these elements dissolved by the acid. Since the same clay minerals are present in all samples, and only some of them contained carbonate and therefore were leached, any substantial effect of the acid on the clay mineral would be revealed also by the analytical data.

Following the treatment with H_2O_2 and HCl, the samples were washed with distilled water, using Pasteur-Chamberlin filter cones as the filtering medium, until no chlorine showed in the wash water.

The samples were then dispersed in distilled water, using $NH_4(OH)$ as the dispersing agent, and fractions coarser than 2 microns, 2 to 1 microns, and minus 1 micron were separated by settling, using Stoke's law to compute settling time. The minus 1 micron fraction was further separated into 1.0 to 0.1 micron and minus 0.1 micron fractions by means of the Sharple's supercentrifuge. Fractionation was repeated until substantially complete separation of the size fractions was attained. The fractions in each case were evaporated over a steam bath to dryness. Ammonia was used as the dispersing agent so that no salt would develop in the fractions to complicate the diagnostic clay mineral data.

Optical data were obtained on all fractions by the usual immersion methods. Differential



LOCATION OF SAMPLES FROM THE FLOOR OF THE GULF OF CALIFORNIA

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TABLE 1.—*Samples from the Pacific Ocean*

Sample No.	Latitude N.	Longitude W.	Depth of water fathoms	Depth in core inches	General location	Character of sample
C 1	32°55.7′	117°36.8′	488	2–7 93–98	San Diego trough	Green silty clay " " "
C 4	33°03.3′	117°25.5′	275	3–8 43–48	San Diego trough	Green silty clay " " "
C 5	33°05.8′	117°21.7′	51	2–5 19–21	So. Calif. shelf	Gray silty sand " " "
C 9	33°14′	117°42.5′	446	0–3 11–13	San Diego trough	Green silty sand Gray silty sand
C 10	33°15.5′	117°36′	249	2–5 35–37	Slope into San Diego trough	Green sandy silt Green silty sand
C 11	33°18.5′	117°34′	36	0–2 10–15	So. Calif. shelf	Gray silty sand " " "
C 12	33°34.4′	117°52′	33	4–6 18–21	So. Calif. shelf	Gray silty sand " " "
C 14	33°30′	118°01′	303	2–6 42–46	Slope into San Diego trough	Gray clayey silt " " "
C 16	33°24.5′	118°11.5′	376	5–10 30–35 100–105	San Pedro basin	Green silty clay " " " " " "
C 18	33°29.5′	118°30.2′	476	2–5 24–28	San Pedro basin	Green silty sand " " "
C 29	33°40.5′	119°16′	457	6–10	NW. of Santa Barbara	Brown sandy silt
C 30B	33°40.7′	119°20′	645	30–32	Santa Cruz basin	Green clayey silt
C 31	33°41′	119°27.5′	1072	75–78	Santa Cruz basin	Green silty clay
F 68	31°36′	116°58′	678	3–8 68–73 99–104	Basin SW. of Todos Santos Bay, Lower Calif.	Green clayey mud Green silty mud " " "
F 72	30°25′	118°32.5′	1640	0–2 8–11 47–50 59–62 83–86 92–94	Abyssal ocean SW of Ensenado Lower Calif.	Brownish-red clay " " " Grayish-red clay Gray-green clayey mud Gray clayey mud " " "
F 84A	27°32′	116°10′	1831	3–6 48–51 69–72	Abyssal ocean SW of Cedros Is., Lower Calif.	Green clayey mud " " " " " "

TABLE 1—*Continued*

Sample No.	Latitude N.	Longitude W.	Depth of water fathoms	Depth in core inches	General location	Character of sample
F 88	28°22.5′	114°34.5′	55	3–7	Sebastian Viscaino Bay, Lower Calif.	Green silty sand
				15–20		Gray green sandy silt
F 89	29°09′	115°58.5′	1031	2–7	Basin SW. of San Antonio Point, Lower Calif.	Green clayey mud
				57–62		" " "
149	32°34.6′	117°27.8′	645	12–15	San Diego trough	Green silty mud
				75–78		" " "
154	32°57′	119°43′	810	0–3	Tanner basin	Green silty clay
				13–18		" " "
				58–63		" " "
				63–68		" " "
				68–78		" " "
163	33°13.3′	120°21.2′	575	1–5	45 mi. W. of San Nicolas Is.	Light-green sandy silt
				5–10		Light-green sandy silt
				20–25		Light-green silty mud
166	33°01.7′	120°34.2′	1717	1–6	Continental slope	
				61–66		
186	30°41′	121°46′	2231	0–3	Red clay area, Abyssal ocean	Brownish-red clay
				27–29		" " "
				67–69		" " "
				81–83		" " "
				87–89		" " "
195	34°11.3′	119°55.5′	290	15–20	Santa Barbara basin	Green silty clay
				60–65		" " "
				100–105		" " "
240	36°39.6′	122°06.2′	1036	2–7	Monterey canyon	Green silty clay
				47–52		" " "
244	36°45.2′	122°02.6′	460	1–5	Monterey canyon	Gray silty sand
				45–50		" " "
				88–90		" " "
253	34°49′	121°10′	292	½–4	25 mi. W. of Pt. Sal	Green clayey silt
				42–47		" " "
				57–62		" " "
258	34°14′	120°02.5′	310	15–20	Santa Barbara basin	Green silty clay
				60–65		" " "
				100–105		" " "
271	32°11.6′	118°19′	1011	2–7	Basin NNW. of Sixty Mile bank	Green silty clay
				17–22		" " "
				82–87		" " "

TABLE 1—*Concluded*

Sample No.	Latitude N.	Longitude W.	Depth of water fathoms	Depth in core inches	General location	Character of sample
284	32°58′	118°52′	881	0–5	San Nicolas basin	Green clayey silt
				20–25		" " "
				75–78		" " "
286	33°06′	118°20.9′	629	10–15	Catalina basin	Green clayey silt
				40–44		" " "

thermal curves were obtained on all samples for which an adequate quantity was available—about two-thirds of all fractionated samples. X-ray diffraction data were obtained on substantially all samples. In some cases the samples were prepared with glycol to check the montmorillonite identification according to the procedure described by Bradley (1945). The chemical analyses were obtained by the usual methods.

The complexity of the mixtures, which seems to be typical of sediments of the sort here studied, makes desirable some statement of the probably reliability of these various methods employed.

X-ray diffraction, thermal analysis, and optical microscopy all observe crystallizations rather than chemical composition. In the study of finer-grained sediments, optical methods first encounter the limitation that particles are unresolved, and only mean indices and aggregate refringences may be observed. X-ray diffraction from powders is of optimum utility for particles of the order of magnitude of 0.1 micron. For smaller sizes, diffraction lines broaden progressively and intensities are reduced. In addition, comparatively minor imperfections within ultimate grains make such grains effective as aggregates, so that X-ray intensities are sensitive both to size and to degree of order, and each sensitivity is effective in the range of the sediments of this study.

In complex mixtures of the type encountered in this study, differential thermal analysis offers less selectivity in the matter of mineral idetifications than do diffraction methods. The more prominent thermal features relate merely to the expulsion of water from hydration states of various levels of stability. However, it has been the rather surprising experience that whereas coarsely crystalline grains may hold water to slightly higher temperature than do fine grains of the same crystalline phase, grains in the size magnitude of the clay minerals exhibit remarkably constant temperature ranges for discrete instance of dehydration. For example, the endothermal feature in a thermal curve which relates to expulsion of hydroxyl water from a given mineral type tends to appear at nearly the same temperatures whether X-ray diffraction diagrams are normal or of attenuated intensity. It follows therefore that diffraction diagrams afford the best criteria of identity for the clay mineral types, while the thermal analyses afford the best estimate of the actual amount present.

This distinction is of particular significance when one considers the nature of mixing. It has been established that clay minerals can occur as mixed crystallizations (Bradley, 1945; Gruner, 1944; Hendricks and Jefferson, 1938) as well as mechanical mixtures of discrete phases. For simple mechanical mixtures, both diffraction effects and thermal diagrams are simple superpositions of separate effects evidenced in proportionate degree. For the mixed layer crystallizations however, diffraction effects are a separate special phenomenon which is less intense than classical diffraction and has been analyzed in only a few fortuitous instances. In these instances, however, thermal features maintained their separate identity, and it seems probable that such thermal features are actually specific for a layer whether or not its adjacent layers are the same or a different species. Although the temperatures of the thermal reactions are not changed when the minerals occur as interlayered mixtures, the intensity of the reactions appear to decrease (Grim, 1947).

In these ocean-bottom sediments, the identi-

TABLE 2.—*Samples from the Gulf of California*

Sample No.	Latitude N.	Longitude W.	Depth of water fathoms	Depth in core inches	General location	Character of sample
GC 5	22°58′	108°14.5′	1558	6–10 42–46	Between Cape San Lucas and Mazatlan	Greey clayey mud " " "
GC 17	24°–09.5′	109°36.6′	270	Snapper	West side of Gulf 80 mi. N. of entrance	Foraminiferal sand
GC 21	25°22.5′	109°57′	1640	22–26	Center of Gulf, 150 mi. N. of entrance	Green silty mud
GC 23	25°33.8′	109°27.6′	530	17–22	East side of Gulf 150 mi. N. of entrance	Green silty mud
GC 27	26°20.4′	110°46.3′	1336	36–40	Center of Gulf, 100 mi. S. of Guaymas	Green diatomaceous clayey mud
GC 29	26°09.5′	111°13.2′	228	40–45	W. side of Gulf 100 mi. S. of Guaymas	Green clayey mud
GC 30	27°14′	111°10′	951	32–38	Center of Gulf, 40 mi. SW. of Guaymas	Green silty sand
GC 35	27°01.6′	111°47.1′	456	52–56	W. side of Gulf 60 mi. W. of Guaymas	Green clayey mud
GC 38	28°07′	112°10.5′	471	78–82	Center of Gulf, 30 mi. S. of Tiburon Is.	Green diatomaceous silty mud
GC 39	28°17′	112°06′	172	8–12	Center of Gulf, 20 mi. S. of Tiburon Is.	Gray-green shelly sand
GC 41	29°05.8′	112°38.3′	119	6–10 30–33	Just NW. of Tiburon Is.	Green silty shelly sand " " " "
GC 42	29°05.8′	112°38.3′	126	7–11	Between Tiburon and Angel de la Guardia	Brownish-gray silty clay
GC 43	29°09.5′	113°02.2′	257	Snapper	SE. of Angel de la Guardia Is.	Green clayey mud
GC 44	30°18′	113°13′	67	2–4 32–36	E. side of Gulf 120 mi. SSE. of Colorado R.	Green silty mud " " "
GC 46	29°55.2′	113°54.2′	269	6–10 34–38	W. side of Gulf 120 mi. S. of Colorado R.	Green diatomaceous clayey mud Green clayey mud
GC 47	29°56′	114°11′	195	12–16 60–65	W. side of Gulf 120 mi. S. of Colorado R.	Green diatomaceous clayey mud Green clayey mud
GC 48	30°42′	114°22′	46	50–54	W. side of Gulf 70 mi. S. of Colorado R.	Brown-gray silty clay

TABLE 2—*Continued*

Sample No.	Latitude N.	Longitude W.	Depth of water fathoms	Depth in core inches	General location	Character of sample
GC 49	30°52′	114°04′	45	4–7	Center of Gulf 70 mi. SSE. of Colorado R.	Green silty sand
				20–23		" " "
GC 50	31°03′	113°40′	30	6–10	E. side of Gulf 70 mi. SSE. of Colorado R.	Green silty sand
				46–50		" " "
GC 51	31°14′	114°14′	28	2–4	Center of Gulf 30 mi. SE. of Colorado R.	Green silty sand
				13–15		Green sandy silt
GC 51B	31°04′	114°22′	44	6–10	W. side of Gulf 40 mi. S. of Colorado R.	Brown-gray silty clay
				14–19		" " " "
GC 53	28°46.5′	113°08.2′	716	12–16	W. side of Gulf 20 mi. W. of Tiburon Is.	Green diatomaceous mud
				60–65		Green clayey mud

TABLE 3.—*Samples from Mission Bay and the Colorado River*

Sample No.	General Location	Character of sample
	Mission Bay, San Diego, Calif.	
A	North side, 20 feet from shore; 1.5′ water	Dark-gray clayey silt
B	North bank; at water line	Dark-gray clayey silt
C	North bank; inner edge of marsh; 1′ water	Dark-gray clayey silt
D	Center of bay	Gray silt
	Colorado River	
1	Lake Mead Virgin Canyon	Red silty sand
2	Lake Mead Black Canyon	Red silty sand
3	Lake Mead Boulder Canyon	Red silty sand
4	River bank, Little Colorado River	Red sandy silt

fication of a phase from its diffraction diagram refers specifically to the presence of grains in which a substantial number of mono-mineral layers are contiguous in the grain. Any possible minor proportion of mixed structures simply have afforded no diffraction data, but have presumably made their contribution to the thermal effects.

DISCUSSION OF NOMENCLATURE

No misunderstanding will arise from the use of illite and montmorillonite as group names inasmuch as no crystallographic classifications within these groups have been made. The kaolinite group minerals include species which are differentiated on the basis of the arrangement of stacking of kaolinite layers, but in the sediments studied, no particles are sufficiently regularly developed to afford specific identification. Diffraction diagrams are more comparable with those afforded by halloysite, but because the name halloysite is associated with a strikingly fibrous mineral, it has rather arbitrarily been decided to refer to the mineral in these sediments as a kaolinite group mineral, or more simply, kaolinite.

The observation in some samples of a chlorite group mineral imposes an additional limitation upon the accuracy of estimations of relative abundance. The crystallization schemes of the chlorite and kaolinite groups are so closely re-

lated that in mixtures of this complexity differentiation is based merely upon the presence or absence of weak additional diffraction lines in the 001 sequence. Thus the probable lower limit of recognition of a chlorite mineral in these mixtures is high, and the possibility always exists that diffraction effects attributed to small amounts of kaolinite may have been occasioned by commensurate proportion of chlorite. This kaolinite-chlorite ambiguity is aggravated by the co-occurrence of varying amounts of montmorillonite, which affords a strong diffraction ring at an angle quite close to that of the first chlorite order.

In Plate 3, figures 2 and 3 are reproduced a few of the diagrams which seem to be unambiguous. Not all of the sediments lend themselves to so clear delineation. The very slight angular resolution between the two low-angle diffraction maxima affords ample opportunity to overlook substantial proportions of the chlorite.

In characterizing this material as chloritic, we are dependent solely on the 001 diffraction sequence. The common macrocrystalline chlorites are characteristically larger in the a and b directions than montmorillonite and illites, but the chlorites of these sediments exhibit substantially the same dimensions as do the other clay minerals.

Nonclay Minerals in the Clay-Size Grades

The separation of the predominantly clay fractions (1 to 0.1 micron and minus 0.1 micron) does not, of course, insure that these fractions are free of nonclay minerals.

In general, X-ray diffraction data indicate that quartz is present in every sample. The 1 to 0.1 micron fractions include usually 5–10 per cent of free α-quartz, and the minus 0.1 micron fractions contain up to 1 or 2 per cent. The same is often true of clastic sediments in general and no particular significance is attached to it. The amount of quartz is usually too small to be revealed by the differential thermal procedure.

Nearly all of the fractions include a greater or lesser amount of a crystalline phase which contributes a diffraction line around 3.1–3.2A and which is clearly unrelated to any of the clay minerals present. In the coarser fractions feldspars are clearly present, and it is assumed that small amounts must carry through to the finer fractions to contribute this one most intense diffraction line (Pl. 3, fig. 1).

It is fairly common in some older sediments to encounter proportions of siderite sufficient for identification by X-ray and thermal methods. In these bottom sediments the fractionation procedure has been such as to minimize any opportunity to establish the presence of such material. In a great many cases, however, clear exothermal effects are noted in the thermal analyses diagrams at about 550°C. Such an effect has been established as characteristic of finely divided siderite disseminated through an inactive matrix (Cuthbert and Rowland, 1947) and it is considered likely that a certain amount of siderite remains associated with the clays in many of the present fractions.

Pacific Ocean Samples

X-ray Data

The diffraction patterns for all of the samples contain lines for montmorillonite, illite, and kaolinite. In most cases the patterns are clearly less well developed than would be obtained from equally complex mechanical mixtures of well crystallized components. Because of the weak and diffuse patterns, the relative abundance of the clay minerals cannot be estimated with a high degree of accuracy. Montmorillonite and illite are more abundant than kaolinite. Montmorillonite and illite are about equally abundant. Chloritic material can be identified in some of the samples.

No correlation is definitely indicated between variation in relative abundance of the clay minerals and location of sample, depth within core, or fineness of particle size.

Differential Thermal Analyses

General character.—Differential thermal analyses were made for substantially all the size fractions of all of the available samples (Figs. 1–15).

The curves for all of the samples *except* those from very close to the shore (C5, Fig. 3, and C11, Fig. 6) are substantially the same. Even the curves for the deep-sea red clay (186) are

FIGURE 1. ILLUSTRATION OF THE FELDSPAR ACCESSORY

Sample no. F89, 2°–7° with and without ethylene glycol treatment. When a prominent diffraction effect in the range 3.1–3.2 A derives from a clay mineral, it is a basal order from montmorillonite. Treatment with glycol shifts this effect to the approximate position of the quartz line at 3.34 A. The two lines referred to are those marked.

FIGURE 2. ILLUSTRATION OF THE NECESSITY FOR, [and the value of] ORGANIC COMPLEX PREPARATIONS

Note that apparently similar diffraction rings of diameter about 1.4 cm. are in the one case shifted to a single ring characteristic of the montmorillonite-glycol complex, and in the other clearly resolved into a montmorillonite-glycol complex ring and a chlorite ring. Filtered Cu radiation. (a) GC51 13°–15° 1–1 micron. (b) GC51 glycol treated. (c) C18 2°–5° 1–1 micron. (d) C18 glycol treated.

FIGURE 3. ILLUSTRATION OF THE RANGE OF COMPOSITIONAL VARIATION FOR 1–0.1 MICRON FRACTIONS

(a) C-29, 6°–10°; mainly montmorillonite; (b) C-4, 43°–48°; mainly illite; (c) no. 244, 1°–5°; kaolinite unusually plain; (d) and (e) C5, 2°–5° and C 30B, 32°–34°; all three groups readily apparent; (f) approximate scale of angular displacements for the more prominent diagnostic diffraction rings. Filtered Fe radiation.

X-RAY DIFFRACTION DIAGRAMS

(See attached sheet for fig. captions.)

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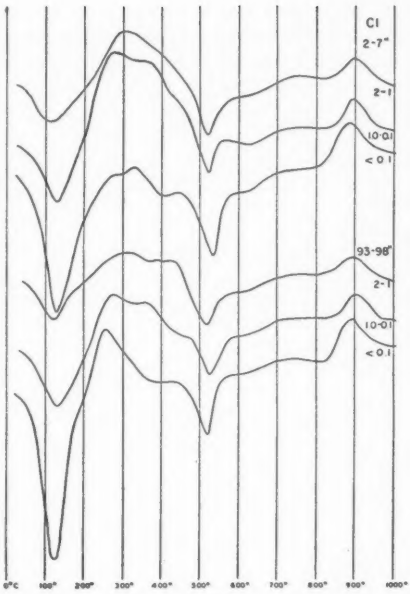


FIGURE 1.—DIFFERENTIAL THERMAL CURVES FOR PACIFIC OCEAN SAMPLE C1

At depths 2″–7″ and 93″–98″ for fractions 2–1, 1.0–0.1 and < 0.1 microns.

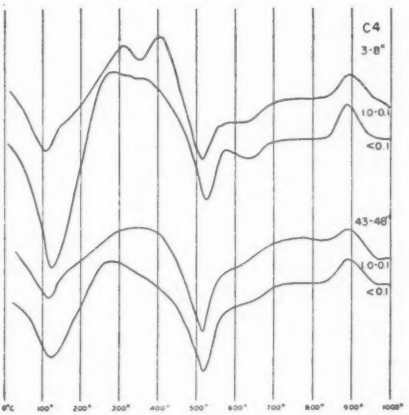


FIGURE 2.—DIFFERENTIAL THERMAL CURVES FOR PACIFIC OCEAN SAMPLE C4

At depths 3″–8″ and 43″–48″ for fractions 1.0–0.1 and <0.1 micron.

substantially the same as those for the other samples.

A comparison of the curves (except those from near-shore samples C5, C11) with those of type clay minerals (Grim and Rowland, 1942)

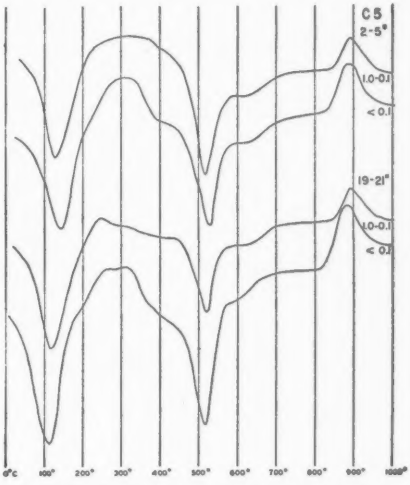


FIGURE 3.—DIFFERENTIAL THERMAL CURVES FOR PACIFIC OCEAN SAMPLE C5

At depths 2″–5″ and 19″–21″ for fractions 1.0–0.1 and <0.1 micron.

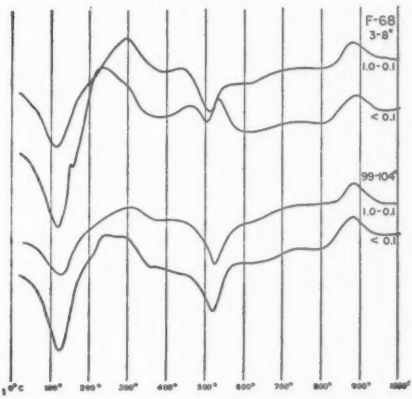


FIGURE 4.—DIFFERENTIAL THERMAL CURVES FOR PACIFIC OCEAN SAMPLE C9

At depths 0″–3″ for fractions 2–1, 1.0–0.1 and <0.1 microns.

and clay mineral mixtures (Grim, 1947) indicate that the chief components are illite and montmorillonite. Generally illite is probably the more important (40% ±), but it is difficult to estimate

the abundance of montmorillonite because the temperature at which certain thermal reactions in montmorillonite take place, notably the reaction for loss of lattice (OH) water, varies with the iron content. For a montmorillonite with relatively high iron content, the temperature of this thermal reaction is about the same as that for illite. As shown later, considerable Fe_2O_3 is carried by a clay mineral component so that the endothermic reaction at 500–600°C. is probably due to both illite and iron-rich montmorillonite. The relative abundance of montmorillonite (30%±) is greater than would be estimated from the endothermic reaction at about 700°C. which is usually taken as the indication of montmorillonite.

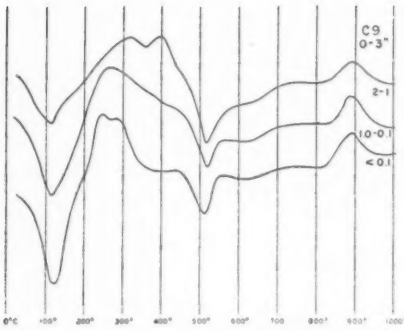


FIGURE 5.—DIFFERENTIAL THERMAL CURVES FOR PACIFIC OCEAN SAMPLE C10

At depths 35″–37″ for fractions 2–1, 1.0–0.1 and <0.1 microns.

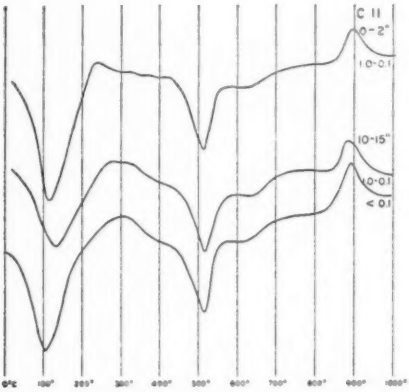


FIGURE 6.—DIFFERENTIAL THERMAL CURVES FOR PACIFIC OCEAN SAMPLE C11

At depths 0″–2″ and 10″–15″ for fractions 1.0–0.1 and <0.1 micron.

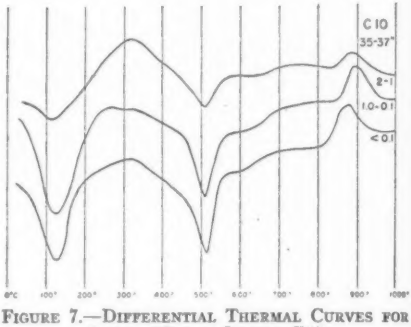


FIGURE 7.—DIFFERENTIAL THERMAL CURVES FOR PACIFIC OCEAN SAMPLE F68

At depths 3″–8″ and 99″–104″ for fractions 1.0–0.1 and <0.1 micron.

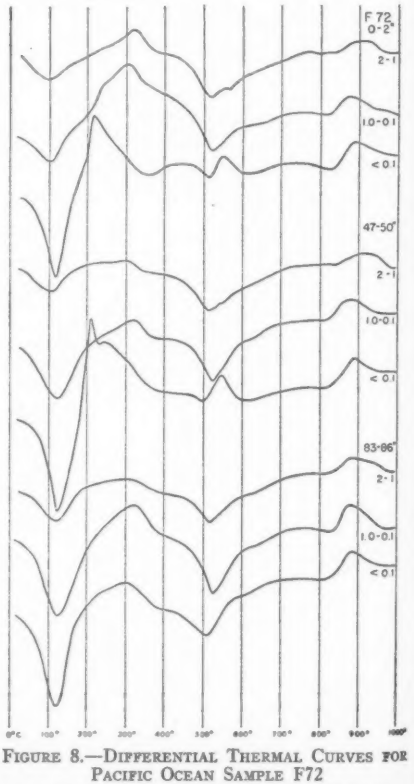


FIGURE 8.—DIFFERENTIAL THERMAL CURVES FOR PACIFIC OCEAN SAMPLE F72

At depths 0″–2″, 47″–50″ and 83″–86″ for fractions 2–1, 1.0–0.1 and <0.1 microns.

All of the curves indicate the presence of kaolinite (20%± in samples obtained some distance from shore). The thermal and X-ray

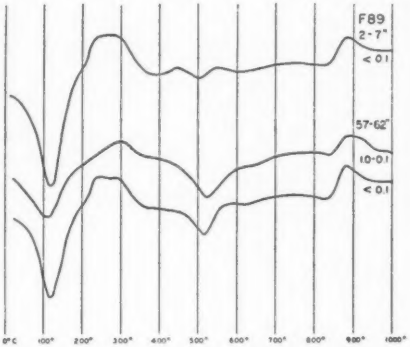


FIGURE 9.—DIFFERENTIAL THERMAL CURVES FOR PACIFIC OCEAN SAMPLE F89

At depths 2″–7″ and 57″–62″ for fractions 1.0–0.1 and <0.1 micron.

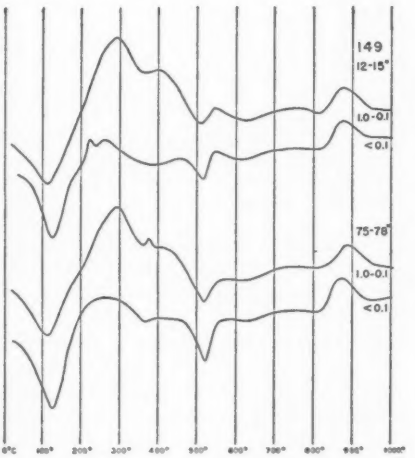


FIGURE 10.—DIFFERENTIAL THERMAL CURVES FOR PACIFIC OCEAN SAMPLE 149

At depths 12″–15″ and 75″–78″ for fractions 1.0–0.1 and <0.1 micron.

data agree in indicating that kaolinite is relatively less abundant than either illite or montmorillonite. The form of the endothermic peak at 500–600°C. and the character of the final exothermic reaction are the features of the curves indicating kaolinite.

The broad exothermic reaction 200–400°C.

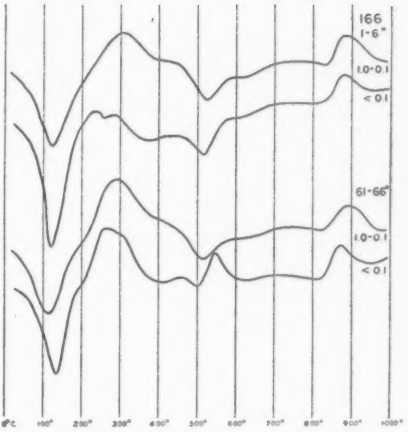


FIGURE 11.—DIFFERENTIAL THERMAL CURVES FOR PACIFIC OCEAN SAMPLE 166

At depths 1″–6″ and 61″–66″ for fractions 1.0–0.1 and <0.1 micron.

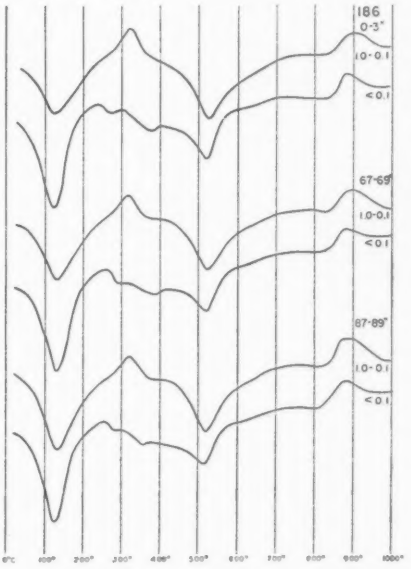


FIGURE 12.—DIFFERENTIAL THERMAL CURVES FOR PACIFIC OCEAN SAMPLE 186

At depths 0″–3″, 67″–69″ and 87″–89″ for fractions 1.0–0.1 and <0.1 micron.

shown by some curves is due to organic material. All of the samples contained as received some organic material, which had to be removed

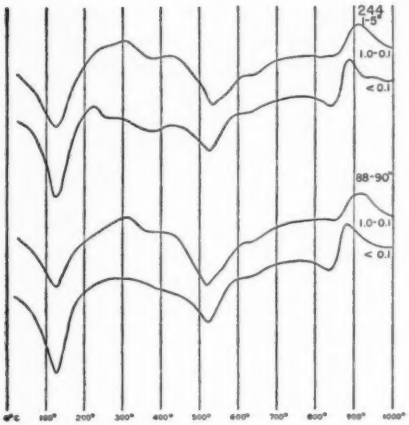


FIGURE 13.—DIFFERENTIAL THERMAL CURVES FOR PACIFIC OCEAN SAMPLE 244

At depths 1″–5″ and 88″–90″ for fractions 1.0–0.1 and <0.1 micron.

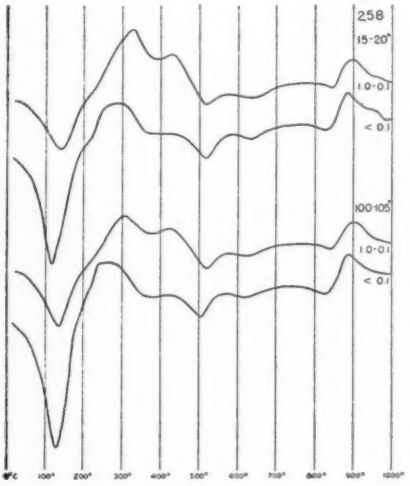


FIGURE 14.—DIFFERENTIAL THERMAL CURVES FOR PACIFIC OCEAN SAMPLE 258

At depths 15″–20″ and 100″–105″ for fractions 1.0–0.1 and <0.1 micron.

by oxidation from most of them in order to obtain X-ray, thermal, and optical analytical data that had diagnostic value. The indications of organic material, therefore, have no definite significance.

The thermal curves generally show slight endothermic reactions at about 300°–400°C. which can be attributed to a ferric iron hydroxide. The size of the reaction is always very small and could account for only a small part

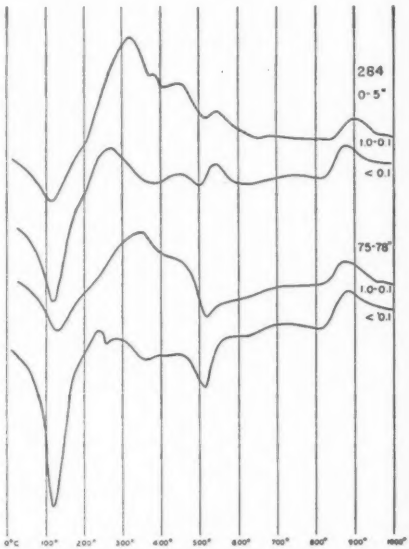


FIGURE 15.—DIFFERENTIAL THERMAL CURVES FOR PACIFIC OCEAN SAMPLE 284

At depths 0″–5″ and 75″–78″ for fractions 1.0–0.1 and <0.1 micron.

of the Fe_2O_3 shown by the chemical analyses (Table 4). It follows that most of the Fe_2O_3 shown by chemical analyses must be present in the clay minerals.

A few samples show at about 565°C. a slight endothermic reaction that is to be correlated with the transition of α to β quartz. The amount of quartz indicated is small (10%) and generally can be detected only in the 2–1 micron size fraction.

Variation in clay mineral content.—A comparison of thermal curves for the series C5–C4–C1 (Figs. 3, 2, 1) and for series C11–C10–C9 (Figs. 6, 5, 4) shows quite definitely that kaolinite is relatively more abundant in the near-shore samples, i.e. C5 (Fig. 3) and C11 (Fig. 6) than in the other samples. There is a progressive decrease in the relative abundance of the kaolinite away from the shore.

In the very near-shore samples, the kaolinite

TABLE 4.—*Chemical analyses*

	166 1–6″ 1.0–0.1	166 61–66″ 1.0–0.1	186 0–3″ 1.0–0.1	186 67–69″ 1.0–0.1	186 67–69″ <0.1	186 87–89″ 1.0–0.1	244 1–5″ 0.1–1.0	244 88–90″ 1.0–0.1	244 88–90″ <0.1	253 ¼–4″ 1.0–0.1	253 57–62″ 1.0–0.1	271 2–7″ 1.0–0.1	271 82–87″ 1.0–0.1
	%	%	%	%	%	%	%	%	%	%	%	%	%
SiO_2	55.65	56.33	52.51	52.41	51.38	52.38	55.83	57.06	54.20	54.46	52.07	53.87	50.98
TiO_2	1.08	1.08	1.03	1.10	.73	1.08	1.00	1.03	.79	1.17	1.17	1.40	1.49
Fe_2O_3	9.00	8.92	10.15	10.61	13.13	10.71	9.13	8.14	9.34	9.56	9.53	9.48	10.98
Al_2O_3	17.85	16.50	19.26	18.84	18.25	18.63	18.19	18.50	19.69	18.37	18.10	18.76	18.85
MgO	3.63	4.28	4.08	4.42	3.43	4.39	4.26	4.43	4.97	3.80	4.49	3.81	4.08
CaO	.91	.94	.83	.82	.82	.84	.55	.62	.18	.79	1.38	.65	1.03
Na_2O	1.49	.74	1.17	.84	.53	.86	.77	.79	.26	.60	.60	.85	.89
K_2O	2.50	2.50	2.74	2.77	2.17	2.54	2.33	2.45	1.97	2.69	2.67	2.74	2.58
L.O.I.	7.69	8.00	7.27	7.42	9.01	7.47	7.65	6.96	9.07	8.20	10.02	7.89	8.77
BaO	N.D.	.65	.67	N.D.	N.D.	.79	N.D.	N.D.	N.D.	none	N.D.	.29	.31
Total	99.80	99.94	99.71	99.23	99.45	99.69	99.71	99.98	100.47	99.64	100.03	99.74	99.96
$-H_2O$ (140°C.)	3.50	4.74	3.92	4.60	4.82	4.74	3.73	3.73	5.38	3.05	3.40	2.88	3.27
SiO_2/R_2O_3	4.01	4.31	3.48	3.49	3.29	3.51	3.98	4.10	3.59	3.78	3.67	3.69	3.28

	Mission Bay D 1.0–0.1	C-1 2–7″ 1.0–0.1	C-1 93–98″ 1.0–0.1	C-4 3–8″ 1.0–0.1	C-4 43–48″ 1.0–0.1	C-14 2–6″ 1.0–0.1	C-14 42–46″ 1.0–0.1
	%	%	%	%	%	%	%
SiO_2	50.87	51.79	52.67	51.48	50.49	52.21	52.29
TiO_2	1.45	1.31	1.33	1.39	1.55	1.32	1.19
Fe_2O_3	8.79	10.09	8.94	9.76	9.56	9.20	9.03
Al_2O_3	22.62	19.72	20.74	20.43	21.41	20.42	21.27
MgO	2.90	3.66	3.27	3.16	3.61	3.36	3.80
CaO	.58	.77	.74	.68	1.23	.83	1.37
Na_2O	.60	.79	.89	.75	1.00	.82	1.21
K_2O	2.39	2.35	2.70	2.20	2.69	2.29	2.77
L.O.I.	9.11	9.74	8.55	10.13	8.39	9.92	7.52
BaO	none	N.D.	N.D.	N.D.	N.D.	N.D.	N.D.
Total	99.31	100.22	99.83	99.98	99.93	100.37	100.45
$-H_2O$(140°C.)	3.60	6.12	5.76	6.22	5.36	6.46	5.30
SiO_2/R_2O_3	3.06	3.37	3.39	3.29	3.12	3.35	3.29

	F-68 3–8″ 1.0–0.1	F-68 68–73″ 1.0–0.1	F-68 99–104″ 1.0–0.1	F-72 8–11″ 1.0–0.1	F-72 59–62″ 1.0–0.1	F-88 3–7″ 1.0–0.1	F-88 15–20″ 1.0–0.1
	%	%	%	%	%	%	%
SiO_2	54.44	53.71	53.93	53.75	53.82	53.53	54.90
TiO_2	1.21	1.40	1.23	1.17	1.15	1.29	1.55
Fe_2O_3	9.62	8.86	9.16	10.12	10.84	8.70	7.58
Al_2O_3	19.15	20.57	20.21	18.95	18.16	18.09	20.19
MgO	2.92	3.69	3.31	3.56	3.44	3.90	3.15
CaO	1.40	.76	1.10	.92	1.21	1.14	.89
Na_2O	1.07	.78	.89	.96	1.00	1.05	.91
K_2O	2.42	2.71	2.72	2.83	2.72	2.91	3.48
L.O.I.	7.75	7.59	7.29	7.05	6.72	9.96	7.68
BaO	N.D.	N.D.	N.D.	.37	.62	N.D.	N.D.
Total	99.98	100.07	99.84	99.68	99.68	100.57	100.33
$-H_2O$(140°C.)	3.60	2.90	3.33	3.50	3.53	3.30	3.03
SiO_2/R_2O_3	3.66	3.47	3.51	3.59	3.65	3.86	3.72

is more abundant in the minus 0.1 micron fraction than in the 1.0 to 0.1 micron fraction. This is strong evidence that the greater abundance of kaolinite in these samples cannot be explained on the basis that most of the kaolinite in the source material is dropped close to shore and other clay minerals are carried farther out. The finer particle size of the kaolinite would suggest that it would be carried to deeper water as well as the other clay minerals.

In the very near-shore sediments, the kaolinite is about as abundant in the deeper part of the cores as in the upper part, indicating that there has been no recent shift to a more kaolinite-rich source material.

The decrease in relative abundance seaward can be explained best by a loss of kaolinite during diagenetic processes. The presence of some kaolinite in the deeper part of the deep-sea red clay core (186) must mean that not all kaolinite is necessarily lost during marine diagenesis and that the transformation of the kaolinite is slow.

The thermal curves suggest that there is slightly more montmorillonite in the near-shore samples than in those farther from shore, suggesting that some montmorillonite also is lost in diagenetic processes.

The data do not indicate other variations in clay mineral composition with variations of depth of water, topography of sea floor, depth in a given core, or particle size of the clay fractions.

Chemical Data

The chemical composition of all the samples are quite similar (Table 4), and the variations are in general of a low order of magnitude. The great uniformity of the composition of the clay in all the Pacific Ocean samples investigated is a striking and significant conclusion of this work.

In the 1.0 to 0.1 micron fraction of the very near-shore samples, the SiO_2/R_2O_3 ratio ranges from 3.12 to 3.39; in the same grade size of the samples taken farther from shore, the ratio is 3.50 to 4 or even higher. In the case of samples 166 and 244, the particularly high SiO_2/R_2O_3 ratio is to be explained by the presence of free silica. Samples 166 and 244 were both taken from localities of steep bottom topography.

With few exceptions, the SiO_2/R_2O_3 ratio is about that anticipated on the basis of the clay mineral composition and does not suggest any appreciable amount of free silica in the colloidal clay fractions of these samples. The difference between the near-shore samples and those from farther off shore is in the higher Al_2O_3 and lower SiO_2 content of the latter. The Fe_2O_3 content is about the same throughout. This variation in chemical composition is in accord with the greater relative abundance of kaolinite in sediments close to shore.

The uniformly high iron content (8–10%) of all the clay samples from the Pacific Ocean is striking. The thermal data have shown that most of this iron is in the clay mineral components, and it is probably distributed in both the illite and montmorillonite. The highest Fe_2O_3 content encountered is in the minus 0.1 micron fraction of the deep-sea red clay. The 1–0.1 fraction of the red clay has about the same Fe_2O_3 content as the same size fraction of the other samples. It should be noted that very small amounts of Fe_2O_3 or the hydroxide are adequate to color sediments red. There seems to be no correlation of Fe_2O_3 content with depth of water, topography of bottom, distance from shore, or depth within core.

The K_2O content of the samples suggests that illite makes up about 40 per cent of the total composition. In the near-shore samples (C5, C11, F68, F88), the amount of K_2O increases with depth within the cores, suggesting that potash is taken up during diagenesis with the formation of illite. There is no regular variation in the K_2O content of the other samples. Probably, at some distances from shore, deposition is so slow that the development of illite keeps up with accumulation and no variation in K_2O or illite content would be expected. The increase in K_2O content with depth in the near-shore sediments is very small indicating that the pickup of K_2O is small, rapid, or both.

The two samples of minus 0.1 micron size grades that were chemically analyzed contain less potash than the 1 to 0.1 micron size grades. This suggests slightly less illite in the finer fraction which is not in accord with the thermal data. The explanation is not known.

The amount of MgO shown by the samples is quite characteristic of sediments composed of

montmorillonite and/or illite. In general the amount of MgO is slightly greater in the samples farther from shore, and in these off shore samples the MgO content increases slightly with depth within the core, thereby suggesting that MgO is taken up very slowly during diagenesis in these samples.

The TiO_2 content is about the same as that frequently found in sediments. There appears no correlation between the variation of TiO_2 and the occurrence of the samples.

The presence of BaO is of interest. The size of the sample available would not permit the analysis for BaO in each sample.

The variation in the values for "loss on ignition" is probably due largely to variations in amount of organic material which was not removed in the H_2O_2 treatment of the samples, and therefore has no significance.

Electron Micrographs

Electron micrographs of several samples (Pl. 4) show the presence of equidimensional units. The electron microscope has shown halloysite, some of the magnesium-rich clay minerals, and some iron-rich montmorillonites to be composed of elongate lath or fiber shaped units (Shaw, 1942), but none of these components are revealed in the Pacific Ocean samples investigated.

Optical Data

Optical values of all of the clay colloid fractions of the Pacific Ocean samples were as follows:

$$\gamma = 1.565 \text{ to } 1.580$$
$$\gamma - \alpha = .020 \text{ to } .025$$
$$(-);\ 2V \text{ small}$$

It was impossible to measure the optical constants of aggregates of mixtures such as compose these samples with sufficient accuracy to detect the small differences that would result from the small variations in clay mineral composition indicated by the other data.

The optical data are those that would be expected for mixtures of illite and montmorillonite together with somewhat smaller amounts of kaolinite. The aggregates frequently appear as a single species, and except for the lack of constancy in optical values and the poor quality of interference figures, might be mistaken for a single mineral. The homogeneity of the aggregates indicates mixing of very small units, and the crude interference figures indicate alignment parallel to the c-axis with little orientation parallel to the a and b axes.

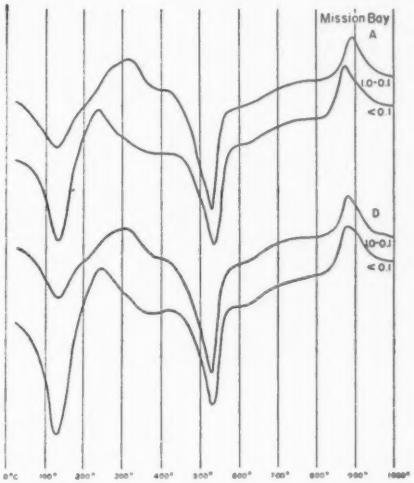


FIGURE 16.—DIFFERENTIAL THERMAL CURVES FOR MISSION BAY SAMPLES A AND D
For fractions 1.0–0.1 and <0.1 micron.

MISSION BAY SAMPLES

X-ray Data

Lines for illite, montmorillonite, and kaolinite are present in the diffraction diagrams of all the samples. The patterns are very poor so that accurate quantitative estimates are impossible. However, it is clear that the relative abundance of each of the three clay minerals is about the same.

Differential Thermal Analyses

The curves for all the samples are about the same and like those shown in Figure 16. The curves indicate the presence of illite, montmorillonite, and kaolinite in about equal relative abundance.

A comparison of the thermal curves for the Mission Bay samples with those from the Pacific Ocean (Figs. 1–15) show that the composition of the Mission Bay clay is like that of the very near-shore samples (C5, Fig. 3) rather than that of the samples from farther off shore. The relatively higher amount of kaolinite shown by the very near-shore sediments is shown also by the Mission Bay sample.

Chemical Data

The 1.0 to 0.1 micron fraction of a Mission Bay sample has a slightly lower SiO_2 to R_2O_3 ratio than the very near-shore Pacific Ocean samples and a considerably lower ratio than those farther off shore (Table 4). This suggests that the kaolinite content of the Mission Bay clay is at least as large and perhaps slightly larger than the very near-shore samples.

The Fe_2O_3 content of the Mission Bay sample is about the same as that in the Pacific Ocean clays.

The MgO content of the Mission Bay sample is slightly lower than that in the very near-shore samples, and considerably lower than that in the clays from farther off shore. The K_2O content of the Mission Bay sample is about the same as that in the near-shore clays and slightly lower than that in the samples from farther off shore.

Since the Mission Bay samples represent the composition of at least some of the source material of the Pacific Ocean clays off the coast of southern California, these data further support the conclusion reached by the study of the Pacific Ocean samples themselves: in processes of diagenesis in this area, kaolinite is lost, K_2O is taken up probably with the formation of illite, and MgO is taken up.

Optical Data

The optical values for the Mission Bay samples are within the same limits as those given for the Pacific Ocean samples.

Gulf of California Samples

X-ray Data

X-ray diffraction data indicate the presence of illite, montmorillonite, and kaolinite in all samples. The diffraction patterns are poor so that estimates of relative abundance cannot be made with a high degree of accuracy. However, the X-ray data do suggest that the relative abundance of the clay minerals in sample G.C.5

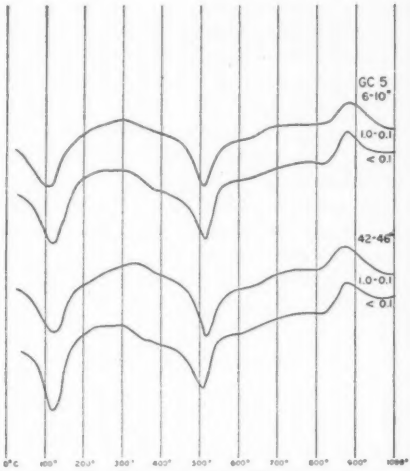


Figure 17.—Differential Thermal Curves of Gulf of California Sample GC5

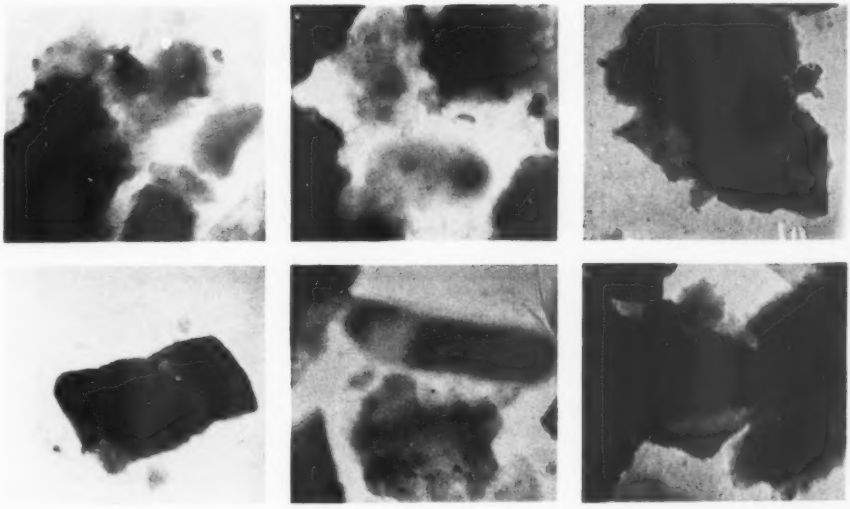
At depths 6″–10″ and 42″–46″ for fractions 1.0–0.1 and <0.1 micron.

from about the mouth of the Gulf is about the same as that in the Pacific Ocean samples; montmorillonite and illite are about equally abundant with kaolinite somewhat less abundant. In the samples from up in the Gulf, the X-ray data suggest that kaolinite is slightly more abundant, being equal or nearly so to the illite and montmorillonite. Chloritic material can be identified in a few of the samples.

Differential Thermal Analyses

Differential thermal curves were obtained for nearly all the samples (Figs. 17–21). In general the curves are similar to those of samples from the Pacific Ocean, Figures 1 to 15. They indicate the presence of illite, montmorillonite, and kaolinite and suggest that illite is generally the most abundant, and kaolinite the least abundant.

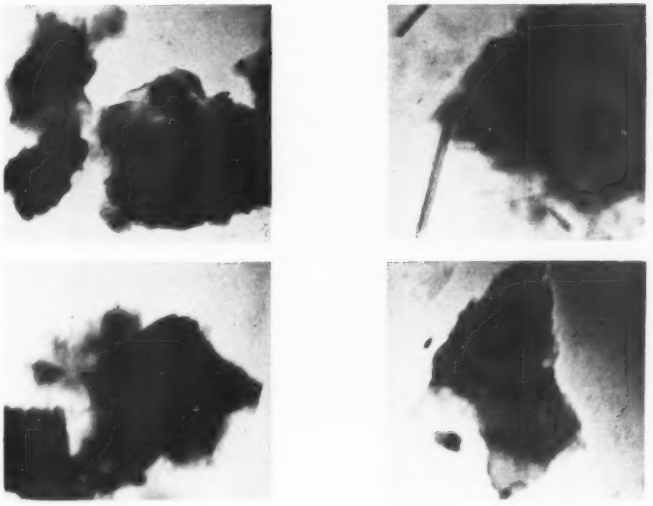
The curves for all of the samples are almost the same. The thermal data do not indicate



C-1
2-7"
1.0-0.1

C-5
19-21"
1.0-0.1

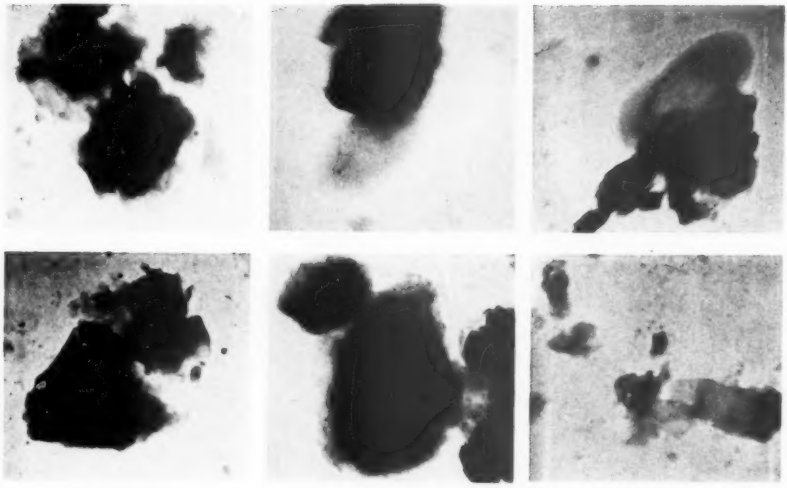
244
88-90"
1.0-0.1



186
87-89"
1.0-0.1

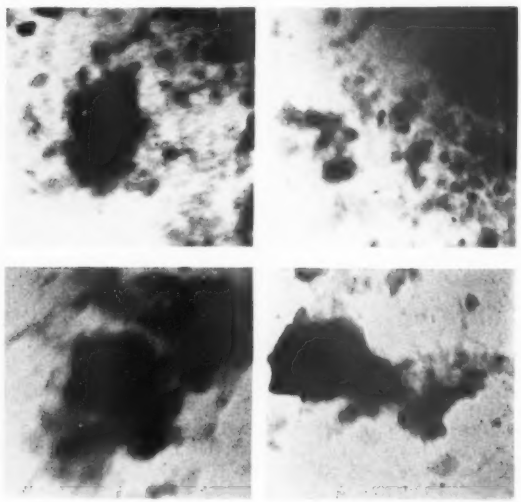
186
87-89"
<0.1

ELECTRON MICROGRAPHS OF PACIFIC OCEAN SAMPLES
23,000X, for samples and fractions indicated



GC 44 2-4" 1.0-0.1 | GC 44 32-36" 1.0-0.1 | GC 5 42-46" <0.1

FIGURE 1. GULF OF CALIFORNIA SAMPLE
23,000X, for samples and fractions indicated.



<0.1 | 1.0-0.1

FIGURE 2. COLORADO RIVER SAMPLES
23,000X, for samples and fractions indicated.

ELECTRON MICROGRAPHS OF GULF OF CALIFORNIA AND COLORADO RIVER SAMPLES

definitely any differences between the relative abundance of kaolinite in the samples from up in the Gulf and sample C5 at the mouth of the Gulf or with the Pacific Ocean samples. The

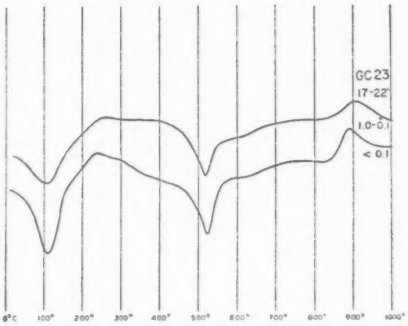


FIGURE 18.—DIFFERENTIAL THERMAL CURVES OF GULF OF CALIFORNIA SAMPLE GC23
At depths 17″–22″ for fractions 1.0–0.1 and <0.1 micron.

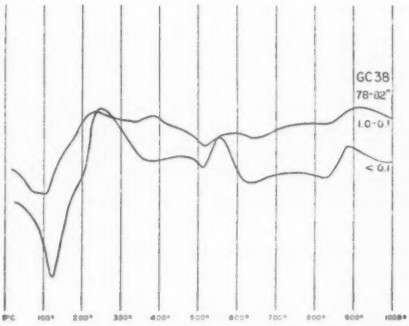


FIGURE 19.—DIFFERENTIAL THERMAL CURVES OF GULF OF CALIFORNIA SAMPLE GC38
At depths 78″–82″ for fractions 1.0–0.1 and <0.1 micron.

data do not confirm the suggestion of the X-ray data that kaolinite is relatively more abundant up in the Gulf, but the accuracy of the estimates of the relative abundance of clay minerals from such thermal curves is low and the data cannot be taken as positive evidence against the suggestion.

In general the thermal curves do not suggest the presence of any hydroxide. The Fe_2O_3 shown

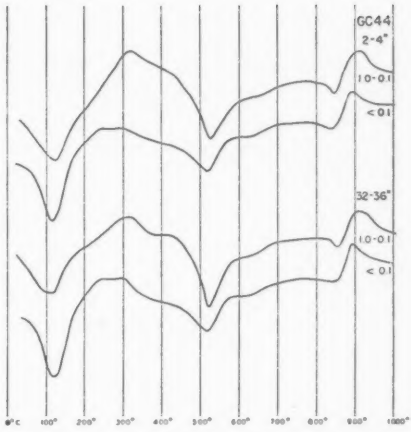


FIGURE 20.—DIFFERENTIAL THERMAL CURVES OF GULF OF CALIFORNIA SAMPLE GC44
At depths 2″–4″ and 32″–36″ for fractions 1.0–0.1 and <0.1 micron.

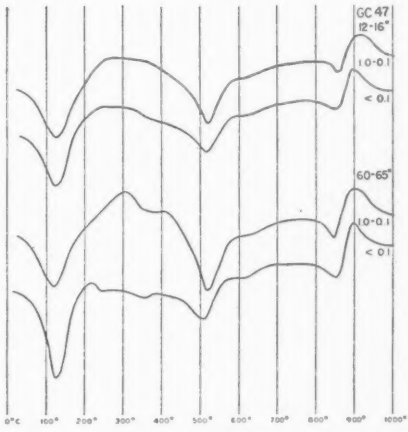


FIGURE 21.—DIFFERENTIAL THERMAL CURVES OF GULF OF CALIFORNIA SAMPLE GC47
At depths 12″–16″ and 60″–65″ for fractions 1.0–0.1 and <0.1 micron.

by the chemical analyses is, therefore, considered to be a component of the clay minerals illite and/or montmorillonite.

The thermal data do not suggest any variation with depth within the cores or any variation between the 1.0 to 0.1 micron and the minus 0.1 micron size fractions.

Chemical Data

A comparison of the analyses for the various samples (Table 5) shows that the SiO_2 to R_2O_3 ratio is quite variable, probably because of differences in the amount of free silica which in some samples is considerably larger than in any similar fraction of a Pacific Ocean sample. Judging from the flood of diatoms in most of the Gulf samples, the free silica of the clay fractions is mainly finely comminuted diatom frustules. The chemical composition of the samples is about what would be anticipated for mixtures of illite, montmorillonite, and kaolinite. The fairly high and variable SiO_2 content makes impossible any worthwhile estimate of the relative abundance of the clay mineral components.

The Fe_2O_3 content is considerably lower than that of the Pacific Ocean samples except in the <0.1 micron fraction of sample GC5 which was taken from about the mouth of the Gulf. The deeper part of the cores generally has slightly more Fe_2O_3 than the upper parts.

The MgO content of the samples from the Gulf is distinctly lower than that of the Pacific Ocean samples. In the deeper part of the cores the amount of MgO is slightly higher than in the upper part, probably additional evidence that MgO is taken up in diagenetic processes at least under some marine conditions.

The K_2O content of the samples is fairly uniform but in general appears slightly higher than that of the Pacific Ocean samples. No correlation is apparent between the K_2O content and location of core sample or depth within the core.

The high content of Na_2O in the 1.0 to 0.1 micron fraction of sample GC5 is without explanation.

The values for "loss on ignition" have no significance because the variations are due to organic material which was not completely removed in the preparation process.

TABLE 5.—*Chemical analyses of samples from Gulf of California and Colorado River*

	GC 5 6–10″ 1.0–0.1	GC 5 42–46″ 1.0–0.1	GC 5 42–46″ <0.1	GC 29 40–45″ 1.0–0.1	GC 38 78–82″ 1.0–0.1	GC 44 2–4″ 1.0–0.1	GC 44 32–36″ 1.0–0.1	GC 47 12–16″ 1.0–0.1	GC 47 60–65″ 1.0–0.1	GC 51 0–4″ 1.0–0.1	GC 51 13–15″ 1.0–0.1	Colo. R. 4 1.0–0.1	Colo. R. 4 <0.1
	%	%	%	%	%	%	%	%	%	%	%	%	%
SiO_2	58.85	57.67	54.40	62.39	66.05	58.25	57.07	59.63	56.56	57.85	57.44	56.37	53.77
TiO_2	.89	.87	.86	.96	.70	.88	1.27	.88	.91	.96	1.00	1.00	.87
Fe_2O_3	5.59	6.34	9.36	4.64	3.86	4.98	5.46	5.78	6.85	6.53	6.59	6.53	6.87
Al_2O_3	19.98	20.13	20.99	16.30	15.76	20.58	21.30	19.70	19.98	20.97	20.94	21.46	23.16
MgO	2.11	2.51	2.30	2.93	2.14	2.61	3.02	3.07	3.55	2.73	3.04	3.05	3.71
CaO	.55	.54	.40	.79	.33	1.21	.45	.21	.64	.37	.58	1.40	.44
Na_2O	1.46	1.29	.39	.93	.85	.68	.51	.56	.52	.53	.49	.46	.17
K_2O	3.43	3.29	1.85	2.18	2.56	3.22	3.50	3.03	2.97	3.34	3.25	2.44	2.60
L.O.I.	6.79	7.22	9.36	8.86	7.63	7.63	7.44	7.33	7.50	6.82	6.94	7.08	8.70
BaO	.15	N.D.	N.D.	N.D.	N.D.	N.D.	N.D.	N.D.	none	N.D.	N.D.	N.D.	N.D.
Total	99.80	99.86	99.91	99.98	99.88	100.04	100.02	100.19	99.48	100.10	100.27	99.79	100.29
$-H_2O$ (140°C.)	3.52	3.80	7.20	5.62	4.95	3.35	3.02	4.92	4.17	4.67	5.20	5.88	4.22
SiO_2/R_2O_3	4.24	4.07	3.42	5.50	6.15	4.16	3.91	4.34	3.94	3.90	3.89	3.72	3.33

Electron Micrographs

The electron micrographs (Pl. 5, fig. 1) do not show the presence of any lath- or fiber-shaped units, indicating the absence of halloysite and any iron-rich or magnesium-rich clay minerals.

Optical Data

Optical values for the 1.0 to 0.1 micron fraction of the Gulf samples were determined as follows:

$\gamma = 1.550$ to 1.565
$\gamma - \alpha = .020$ to $.025$
$(-)$; 2V small

The indices of these samples are generally slightly lower than those of comparable samples from the Pacific Ocean, which is in accord with the slightly lower iron content and perhaps the relatively larger amount of kaolinite in the Gulf samples. The fractions from sample GC5 at the mouth of the Gulf have optical values about like those of the Pacific Ocean samples. The presence of larger amounts of free SiO_2 in the Gulf samples would also lower the indices, but lower indices were found in samples with a low SiO to R_2O_3 ratio indicating that this is not the only explanation.

In general the indices of refraction of the minus 0.1 micron fraction were slightly higher than those of the 1.0 to 0.1 micron fraction. The previous suggestion that the Fe_2O_3 content is slightly higher in the minus 0.1 fraction is a possible explanation.

No other correlation of optical data with location of core, particle size, or depth within core could be detected. Again it should be noted that the accuracy of determination of optical values on such aggregates is very low.

As in the Pacific Ocean samples the homogeneity of the aggregates is frequently striking, indicating intimate mixing of very small units with considerable alignment parallel to the C-axis.

Colorado River Samples

X-ray Data

The X-ray data for the 1.0 to 0.1 micron fractions of the Colorado River samples indicate the presence of illite, montmorillonite, and kaolinite, and suggest that they are present in about equal abundance. The data for the minus 0.1 micron fractions indicate the presence of the same three clay minerals, but the patterns are too poor for any estimation of relative abundance.

Differential Thermal Analyses

All of the samples showed about the same thermal curves and are like the examples shown in Figure 22.

The curves indicate mixtures of illite, kaolinite, and montmorillonite. The amount of kaolinite, particularly in the 1.0 to 0.1 micron fractions, is considerably larger in the Colorado River samples than in those from the Gulf of California which suggests the loss of some kaolinite during diagenesis in the Gulf.

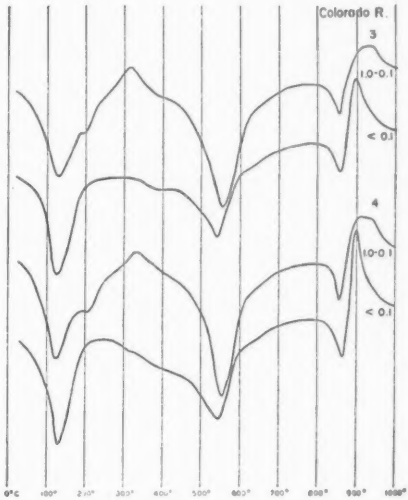


FIGURE 22.—DIFFERENTIAL THERMAL CURVES OF COLORADO RIVER SAMPLES 3 AND 4
For fractions 1.0–0.1 and <0.1 micron.

The thermal curves for the Colorado River samples, unlike those from either the Gulf or the Pacific Ocean, show definite differences between the 1.0 to 0.1 micron and the minus 0.1 micron fractions, principally in the relatively larger amounts of kaolinite in the coarser fraction. It may be concluded that the keolinite in the river samples is mechanically mixed with the other clay minerals rather than intimately interlayered with them. It seems, therefore, that at least some of the intimate interlayering shown by the Ocean and Gulf samples is a consequence of diagenetic processes.

The thermal data do not suggest a relatively larger amount of montmorillonite in the Colorado River samples than in the Gulf samples.

Chemical Data

Chemical analyses of two size fractions of a Colorado River sample are given in Table 5. In general the chemical composition is similar

to that of the Gulf samples; the SiO_2 to R_2O_3 ratio of the river samples is slightly less, which is probably due to more kaolinite and less free silica.

The Fe_2O_3 content is about the same in the Colorado River and Gulf samples. The MgO content is higher in the river samples than in the Gulf samples, but still somewhat below that of the Pacific Ocean samples. The significance of this variation is not known.

The Colorado River samples contain appreciably less K_2O than the Gulf samples. The Na_2O of the river samples is lower also than that of the marine samples.

Electron Micrographs

Electron micrographs of the Colorado River samples (Pl. 5, fig. 2) do not indicate the presence of any lath- or fiber-shaped units.

Optical Data

The optical values of the Colorado River samples are the same as those of the Gulf samples.

Discussion and Conclusions

Data from the Pacific Ocean and Mission Bay samples and the Gulf of California and Colorado River samples indicate that kaolinite is lost during diagenesis under marine conditions. The data suggest further that the loss is slow so that even after fairly long periods of time not all the kaolinite is destroyed. If this conclusion is generally applicable, sediments deposited under marine conditions should be composed largely of clay minerals other than kaolinite. Clay mineral investigations of recent and ancient sediments are too scant to determine the general validity of this conclusion; however, the writers have been impressed by the fact that in the many definitely marine samples which they had studied and in others reported in the literature, illite and perhaps montmorillonite rather than kaolinite are the dominant components. In paleogeographic interpretations of an ancient sediment, an abundance of kaolinite in a sediment would suggest that it was not deposited under marine conditions.

The product of the diagenetic alteration of kaolinite is probably a mica type clay mineral, i.e., illite or perhaps the chloritic clay mineral to be noted presently.

It has long been considered that potash is taken up by the sediments of the oceans, and the present investigation affords supporting evidence.

Illite occurs abundantly in surficial material that is undergoing alteration due to weathering processes. If weathering is not very intense or long continued, a degraded illite weathering product is formed as a result of the partial removal of potassium from between the unit layers of the mineral. Long and intense weathering may alter the illite to montmorillonite, kaolinite, or some other material depending on the conditions of weathering. The clay debris carried to the sea in many areas undoubtedly contains much degraded illite. Such material would rapidly pick up potassium and the illite would be built up again. It is likely that such material does the major amount of picking up of potassium from ocean waters. Some potassium might also be adsorbed by the montmorillonite, but the degraded illite would probably be the major factor because of its great avidity for potassium.

The degraded illite carried to the sea would certainly have a variable K_2O content because of variations in time and conditions of weathering in different land areas. One would not expect that adsorption of potassium from ocean waters would equalize such differences completely, and hence the K_2O content of sediments from different areas of the ocean would be expected to show variations. This would explain the higher content of K_2O in the sediments of the Gulf of California than in those of the neighboring part of the Pacific Ocean.

The data afford considerable evidence that magnesium is taken up slowly by marine sediments during diagenesis. Magnesium is generally a component of illite which may also be lost during weathering. Degraded illite, therefore, might be expected to pick up some magnesium. Magnesium may also enter into base exchange reactions; its relative exchangeability is such that it would tend to be adsorbed in this way by montmorillonite or illite.

It seems likely that the chloritic material

which could only be identified definitely in a few samples perhaps because of the character of the analytical data may well be a diagenetic product whose development is tied to the tendency of marine sediments to take up magnesium. The study of many marine sediments in the laboratory of the Illinois State Geological Survey has revealed the presence of a chloritic clay mineral frequently intimately intergrown with the illite. Future work may reveal that such chloritic material is quite significant as a paleogeographic criterion.

A surprising discovery was the widespread occurrence of montmorillonite in the sediments investigated. In general very little montmorillonite seems to be lost during marine diagenesis. The study of ancient marine sediments has generally revealed little montmorillonite and it seems likely that montmorillonite is lost by metamorphism rather than by diagenesis.

The high content of Fe_2O_3 in the Pacific Ocean sediments is striking. Its source is unknown and the available data do not reveal whether or not it is unique. Ancient sediments with which the writers are familiar do not in general have such a high iron content. In fact the lower content of Fe_2O_3 in the Gulf of California sediments suggest that it is a rather local phenomenon.

The low degree of crystallinity of the clay minerals, their very small size, and their intimate intergrowth is noteworthy. In general in ancient sediments, the crystallinity is of a higher order and the individual clay minerals are more discrete units. For example, particle size fractionation of ancient sediments will frequently separate to some degree the clay mineral components. The analytical data, particularly the thermal curves, show that this is not the case for the marine samples investigated. The cause of this very intimate mixing is undoubtedly to be correlated with the diagenetic changes the sediments are undergoing. It seems likely that when more is known of diagenesis and recent sediments, the crystallinity and intimacy of mixing of clay minerals may become an important paleogeographic key.

No glauconite could be identified specifically in any of the samples studied. Recent researches (Hendricks and Ross, 1941) have shown the similarity of glauconite to the illite group of clay minerals. The high iron content is possibly to be explained by the presence of some glauconite. Information regarding the illite clay minerals and glauconite is not sufficient to permit their clear differentiation in materials of the kind investigated.

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MANUSCRIPT RECEIVED BY THE SECRETARY OF THE SOCIETY, NOVEMBER 29, 1948

PROJECT GRANT 268-39.